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PROTEROZOIC (?) ROCKS IN UTAH

BY A. J. EARDLEY AND R. A. HATCH

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ABSTRACT

New sequences of supposedly Proterozoic strata are described on Promontory Point, Fremont Island, Antelope Island, Sheeprock Range, and the northern Wasatch Mountains. The Big Cottonwood Canyon sequence of Walcott is emended, and a fragmentary sequence in the west end of the Uintas is given. These and other sections already described are correlated insofar as possible and are related to known Cambrian strata overlying them and to the structural features of the region.

The rocks were deposited in deep troughs in and between the northern Utah highland and the Colorado Plateau highland. Each trough was filled with sediments more than 12,000 feet thick. The beds are about half argillaceous and half arenaceous.

The Fremont Island sequence contains a tillite 700 feet thick, and one 1100 feet thick occurs in the Sheeprock Range. Other tillites in the same and in other sections indicate repeated ice advances. The interbedded tillites and associated laminated argillites lead speculatively to the conclusion that the entire sequence of argillaceous, sandy, and tillitic strata may have been laid down during a glacial epoch.

The age of the strata is either lower Cambrian, Proterozoic, or both. Study to date of the degree of metamorphism of Cambrian and underlying strata leads to no definite conclusion. Some deformation and erosion occurred before sands of Lower (?) Cambrian age were deposited, but in light of the highland and trough theory the angular unconformity around the highland margins may not be of great time significance.

INTRODUCTION

LOCALITIES OF EXPOSURE

Great thicknesses of strata, presumably of Proterozoic age, crop out in the Brigham City-Ogden section of the Wasatch Mountains, in Big Cottonwood, Little Cottonwood, and American Fork canyons of the central Wasatch, in the Santaquin section of the southern Wasatch, in the Sheeprock and Simpson ranges 65 miles southwest of Salt Lake City, and on several of the islands of Great Salt Lake including Promontory Point. The beds forming the core of the Uinta Mountains have also been considered pre-Cambrian by previous investigators, and a sequence at the

west end is included in the following discussion. The localities referred to in the present report are shown on the index map (Fig. 1).

MAJOR DIVISIONS OF PRE-CAMBRIAN ROCKS

The major divisions of pre-Cambrian rocks in north-central Utah are an ancient series of gneisses, schists, metaquartzites, granites, and pegmatites, and a younger much less metamorphosed series whose outcrops have just been listed. The older crystalline rocks are treated in another paper (Eardley and Hatch, 1940). The younger strata may be divided into many lithologic units and consist chiefly of arenaceous and argillaceous sediments, now chiefly quartzites and phyllites. Tillites are common, and thin limestone beds have been found.

PREVIOUS WORK

The pre-Cambrian rocks were first observed by Captain Stansbury in 1849 (1853) and then by King, Emmons, and Hague (1877). In the late seventies Walcott (1886, p. 38-39) visited the Big Cottonwood Canyon area a few years later and published a description in 1886 of the quartzite sequence exposed there. Blackwelder (1910, p. 517-542) studied the thrust structures in the Ogden Canyon area in 1909 and in the resultant paper discussed briefly the pre-Cambrian rocks, principally the quartzites, of the Wasatch Mountains. A review of the work on the Algonkian rocks of north-central Utah is found in Hintze's (1913, p. 92-103) paper, in which he recognizes an angular unconformity between the Cambrian quartzite (Tintic quartzite) and the underlying 10,000+ feet of quartzite and metamorphosed shale. The entire sequence had been placed by King and Walcott in the lower Cambrian. Hintze also discusses at length the "tillite" problem and agrees with Blackwelder (1932, p. 289-304) as to its glacial origin. In this last paper Blackwelder summarizes the previous work and adds to the information on the "ancient glacial formation in Utah."

Butler, Loughlin, and Calkins (Butler, *et al.*, 1920, p. 222, 234, 324) also review briefly the Cambrian and Algonkian problem, following essentially the earlier views of Blackwelder and Hintze. Schneider (1925, p. 35-37) adds some observations to the Cambrian and Algonkian unconformity. In 1930, east of Santaquin, Eardley (1932, p. 311-314) found 500 to 1000 feet of Algonkian quartzite and shale in sedimentary contact on the Archean complex. The Algonkian quartzite and shale beds are in turn overlain unconformably by the Cambrian quartzite.

Powell (1876, p. 137-140) described briefly the pre-Cambrian rocks of the eastern portion of the Uinta Mountains and recognized an Archean complex—the "Red Creek quartzite"—and a thick Algonkian series of quartzites and shales—the "Uinta group." King and Emmons (1877, p. 270-271) originally correlated the "Uinta quartzite" with the Weber

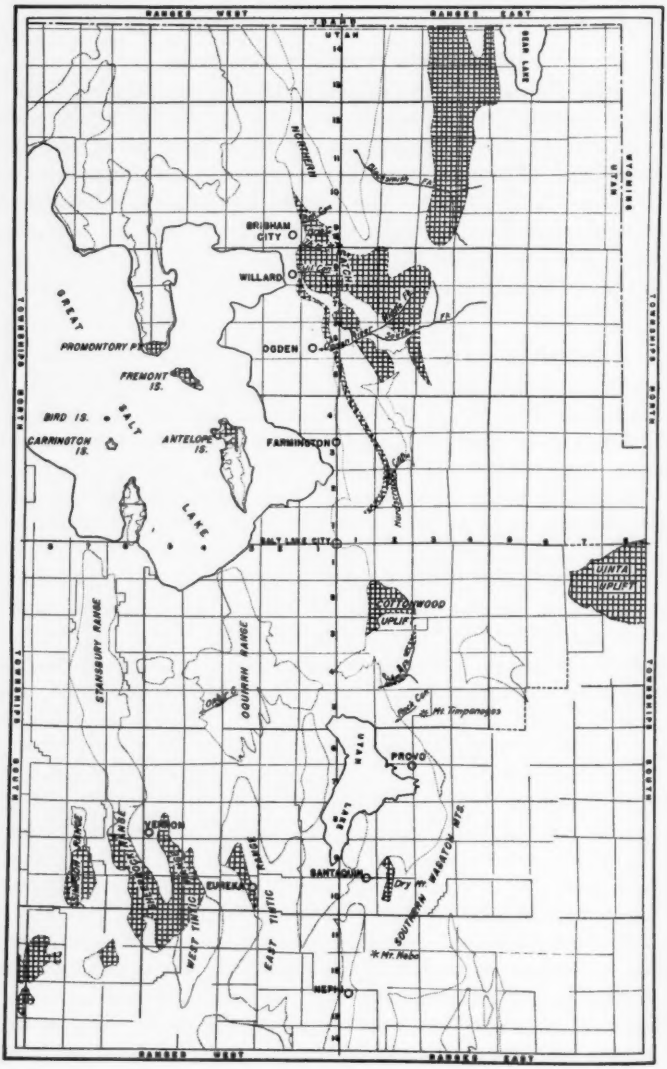

FIGURE 1.—*Index map of north-central Utah*
Showing exposures of Cambrian quartzite and Proterozoic (?) strata

quartzite of Pennsylvanian age, but Emmons (1890, p. 258) later reached the conclusion that it was pre-Cambrian. Weeks (1907, p. 444) considered the Red Creek quartzite Algonkian and the Uinta group Cambrian. Butler (1920, pl. 4) follows Weeks in mapping the Uinta group as Cambrian. The new geologic map of the United States (1935) shows the Uinta group to be Algonkian. The "Uinta series" is considered Algonkian by Forrester (1937, p. 637-638), although a break between the Cambrian beds and "Unita series" was not found.

In a brief summary of the pre-Cambrian rocks of Utah and Wyoming Blackwelder (1935, p. 153-157) recognizes two general divisions, as already well defined, and calls them the "quartzite-slate system" and the "metaquartzite and schist system."

Two papers were published by Hinds (1935, p. 1-52; 1936, p. 53-136) after a regional field review of the Algonkian rocks in western North America. In his first paper he postulates peneplanation of regional extent before and after the Algonkian sediments were deposited, epeirogeny in the ep-Algonkian interval, and a break in sedimentation everywhere in western North America between the Algonkian and Cambrian. In the second paper he divides the Algonkian strata into two groups separated by a considerable time interval during which orogenic deformation, intrusion of granite, and deep erosion occurred. The "Cottonwood series of the Wasatch, Uinta, Oquirrh, and other ranges in northern Utah" is correlated with the older of these two groups and is believed to be older than the Belt series of Montana.

This very brief review reveals considerable confusion in rock unit nomenclature and lithologic terminology as well as correlation of the units in Utah with others in adjoining States. The chief need is additional information.

PRESENT WORK AND ACKNOWLEDGMENTS

Although the rocks presumably of Proterozoic age in north-central Utah have been mentioned in numerous articles, very little systematic work has been done in describing and correlating the exposed sequences. In the present report five thick sections are described for the first time, and one is emended. An attempt is made to define and correlate the major units of the old sedimentary sequences and to decipher their age, origin, and metamorphic history.

The field work was done by Eardley, the petrography by Hatch, and the preparation of the manuscript by Eardley. The authors are indebted to Eliot Blackwelder and T. S. Lovering for criticisms of preliminary manuscripts and to W. F. Hunt for assistance in petrographic interpretations. The field expenses were partially defrayed by a grant from the Horace H. Rackham School of Graduate Studies Research Fund.

ISLAND SEQUENCE

GENERAL CHARACTERISTICS

A large part of a sequence, supposedly of Proterozoic age, is exposed on Promontory Point where it is easily accessible and without structural

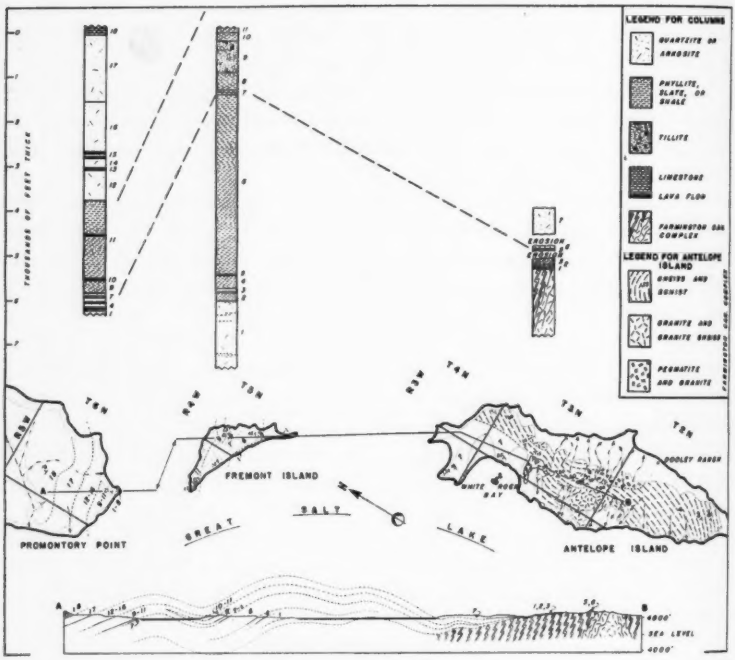

FIGURE 2.—*Island sequence*

Columnar sections, map, and postulated cross section through Promontory Point, Fremont Island, and Antelope Island.

complexities. An equally large part of the same sequence is exposed on Fremont Island, but lack of boats capable of reaching the island make it almost inaccessible. Smaller parts of the section are found on Antelope and Bird islands and on Little Mountain (Fig. 1). The beds consist of phyllites, slates, argillites, shales, quartzites,[1] tillites, arkoses, and lime-

[1] The terms metaquartzite, quartzite, and quartzitic sandstone, as used in this paper, signify a classification of the arenaceous rocks based on the genesis of the textural and structural features in them.

(1) Metaquartzite signifies a relatively pure siliceous rock, once a sandstone, that has been dynamically metamorphosed. Megascopically, a metaquartzite has a fused appearance and usually shows a faint trace of schistosity. Microscopically, the rock has a cataclastic texture if the process of metamorphism has favored crushing, and has a granoblastic texture if metamorphism has reached the

stone. The argillaceous rocks constitute over half the total thickness of beds exposed. The quartzites comprise about one-fourth, and the tillites and limestones about one-fourth.

PROMONTORY POINT

The lowest beds on Promontory Point are exposed at the loading pier where they appear from beneath the sediments of Great Salt Lake. The highest beds are found on the hillside to the north.

The section measured from the middle Cambrian limestone downward is described below and diagrammed in columnar form in Figure 2. The unit numbers of the descriptions correspond to the unit numbers of the column.

Units	Thickness in feet
18. Limestone, dark gray. Middle Cambrian fossils reported	?
17. Quartzite, pink, hard, considerably brecciated, weathers rusty. Apparently conformable on unit 16. (*See* following discussion.)	1500
16. Quartzites, pink, gray, dark gray, pinkish purple, purple, purple red, and tan gray. A few thin greenish shale beds in the quartzites. Quartzite mostly purple near top. Some beds brecciated. Slickensides in several directions. A few ripple-marked surfaces. One pebble conglomerate bed observed.	1100
15. Two 5-foot lava flows (?) separated by 30 feet of pink quartzite. (*See* thin-section description that follows.)	40
14. Quartzite, dark gray, interbedded with greenish quartzitic shales. Most all bedding plane surfaces exhibit micaceous green and gray aspects. Beds are slightly wavy and have dark- and light-gray bands ½ to 1½ inches thick. Surfaces commonly have mottled bumps ⅛ to ½ inch in diameter.	250
13. Lava flow. (*See* thin-section description that follows.)	5
12. Quartzite. Same as unit 14. Conformable on unit 11	700
11. Phyllites, and minor amount of slate, dark gray. Some beds hard and exhibit cleavage at angle to bedding planes in contrast to foliation generally parallel with bedding. Certain beds split almost paper thin, and others break down into smooth powder. The phyllites are badly disturbed in upper 1000 feet, and the minor structures show thrusting to the northwest. An 18-inch quartzite bed is 100 feet from top. A thin lava flow is 750 feet from top.	1750
10. Lava flows. (*See* thin-section description that follows.)	20
9. Phyllites and some slate, light gray, mirror-like reflection from slabs in sun, and in field called "silver phyllites." Major part of unit is made up of laminations 1 to 3 millimeters thick. Some of the phyllites border on schists in appearance.	300
8. Sandstone, gray, calcareous and quartzitic	10

stage of recrystallization. All grades exist between rocks showing the cataclastic texture and those showing the granoblastic texture. Granulated zones and elongated grains are responsible for the schistose structure shown by this type of rock.

(2) Quartzite signifies a sandstone that is cemented by silica and has not been perceptibly metamorphosed. Megascopically, it is a dense rock with low porosity. It breaks across most of the grains when fractured, giving the fresh surfaces a glassy appearance. Microscopically, the outlines of the original quartz grains are preserved, and secondary growth of quartz around the original grains may be distinguished.

(3) Quartzitic sandstone differs from a quartzite chiefly in having greater porosity and a less glassy appearance. It is midway in appearance between a slightly friable sandstone and a quartzite. Fractures pass around the majority of the grains and not through them. Many gradations between quartzitic sandstone and quartzite have been observed in the rocks referred to in this report, and their distinctions are necessarily arbitrary.

Units	Thickness in feet
7. Limestone, dark gray. Beds 1 to 6 inches thick. Laminations 1/16 to 1/4 inch. Ankerite in veinlets. (*See* following description.)	5
6. Quartzites, pink and gray. Brecciated in certain beds	75
5. Quartzite, dark gray	2
4. Sandstone, tan gray, quartzitic, and shale, green, micaceous. Unit partly covered	100
3. Chert, light pinkish tan. Weathers in small cross ridges. Appears to be unconformable on units 2 and 1	35
2. Quartzite, dark gray. Like unit 5. On weathered surface quartz grains stand out and reveal rounded to subangular outlines	30
1. Quartzite, light gray, grained with quartz particles varying up to 5 millimeters in diameter. Base covered by Great Salt Lake sediments	35
Total exposed thickness	5957±

The beds transitional from units 16 to 17 are mostly covered, but strike and dip readings of both units reveal no angular unconformity between them. There is little or no shale between the quartzite of unit 17 and the overlying limestone of unit 18 as is found in the Wasatch Mountains to the east and the Oquirrhs to the south.

No evidence of an unconformity between rocks dominantly phyllitic—units 9 to 11—to rocks dominantly quartzitic—units 12 to 17—was found.

The presence of lava flows of similar composition in both sequences suggests deposition of both during an episode of volcanism. The units marked as lava flows—nos. 10, 12, 13, and 15—are greatly weathered, greenish-gray rocks that contain phenocrysts of augite 4 to 5 millimeters in length embedded in a fine groundmass. Some plagioclase feldspar phenocrysts also occur. The texture of the groundmass is coarsely trachytic. The association of a highly altered feldspar with abundant augite suggests that the feldspar is andesine or labradorite. In addition to augite and plagioclase, small laths of a pleochroic material in an advanced stage of alteration occur in the groundmass and appear to be biotite. Apatite is also rather abundant.

The microscopic features of these rocks suggest they are andesite or basalt porphyries, but the lithology and field relations are not entirely sufficient to prove whether they are sills or flows. The occurrence of mafic flows in the phyllites of nearby Little Mountain leads the writers to ascribe the mafic porphyries of Promontory Point to extrusive activity also.

FREMONT ISLAND

A calculation from strike and dip readings and a pacing traverse reveals over 5000 feet of dark phyllites, and a 700-foot tillite on Fremont Island. This is the greatest argillaceous section anywhere exposed in the north-central Utah area. A rather dark-gray laminated limestone about 100 feet thick below the tillite is also unusual. The phyllite and tillite beds are underlain by quartzite beds.

Blackwelder (1932, p. 300) surmised from Stansbury's (1853, p. 159, 163) fragmentary description of the rocks on Fremont Island that tillites and laminated phyllites occurred there. It might also have been presumed from Stansbury's writings that the older crystalline complex was also exposed, but such terms as "talcose slate overlain by granite and gneiss" evidently refer to the rocks here called phyllites and the overlying tillite containing aplite and quartzite boulders.

The major units and approximate thicknesses of each as seen and estimated on Fremont Island (Fig. 2) are as follows:

Units	Thickness in feet
11. Phyllite, dark gray, and shiny light-gray varieties. Pronounced slaty cleavage developed parallel to bedding. Thin quartzite layers interbedded.	100+
10. Greenstone, phyllite, and thin tillite layers interbedded. (Description of greenstone follows.)	200
9. Tillite, dark purplish-grey phyllitic matrix. Boulders all sizes up to 3 feet in length. Pebbles and boulders are mostly aplites and quartzites. (Their description follows.)	700
8. Phyllite, dark gray.	300
7. Dolomite, fairly dark gray, laminated. Cut by network of quartz and dolomite veinlets. (Description follows.)	100
6. Phyllites, shiny light-gray and dark-gray varieties. Some talcose and smooth, some granular. Some with scattered staurolite crystals, some with hematite pseudomorphs after pyrite. Slaty cleavage in some beds.	4000
5. Tillite or coarse arkosite. All pebbles of quartz less than ½ inch in diameter	3
4. Phyllite, same as unit 6.	300
3. Chlorite schist. (*See* microscopic description that follows.)	60
2. Phyllite, same as unit 6.	200
1. Quartzite with schistose structure. Arkosites and phyllites probably interbedded. The quartz grains are strongly strained, and biotite flakes along the foliation planes are altering to chlorite and limonite.	1500+
Total exposed thickness	7500±

The dolomite of unit 7 has coarsely crystalline masses probably as irregular veins crosscutting the primary laminae. The magnesium may, therefore, be secondary, and the rock originally a limestone. The dolomite gives a strong test for iron when dissolved in hot HCl. Muscovite is the most abundant of the grains in the insoluble residue, of which it constitutes about 90 per cent. Other minerals present are hornblende, quartz, magnetite, zircon, rutile, leucoxene, and hematite. If the dolomite were originally a limestone, it would be similar to the Promontory Point limestone and Antelope Island limestone, on the basis of its laminations, iron content, and mineral assemblage of the insoluble residue. These limestones are believed to be the best tie between exposed sequences.

The boulders and pebbles in the tillite of unit 9 are, as already mentioned, chiefly aplites and quartzites. One pebble selected in the field as typical of the majority of the boulders and pebbles proved to be composed of 60 per cent quartz, 40 per cent orthoclase, and a little albite. The texture was rather coarse for an aplite. Other pebbles of aplite con-

tained a little muscovite and slightly less quartz (50 per cent). Some pebbles were finer-textured than the pebble selected as most typical. One medium-grained quartzite pebble was white and somewhat friable; another quartzite pebble was black. Its color was due to iron oxide in a little calcite cement. More abundant than these two quartzites is a coarsely crystalline quartzose rock which may be either vein quartz or metaquartzite.[2]

The greenstone of unit 10 and the chlorite schist of unit 3 are much altered, and their original state is difficult to decipher. The greenstone has inconspicuous schistose structure, is composed of 70 per cent green hornblende, 10 per cent chlorite, 10 per cent quartz, and some clinozoisite, epidote, and calcite. The hornblende is about midway in composition between tremolite and actinolite which indicates a ratio of about one Mg to one Fe.

As the hornblende is relatively low in iron, and as only 20 per cent of the minerals of the rock contain aluminum, it seems evident that the original rock was low in these elements. If an impure limestone were altered by magnesium-bearing solutions such as those that dolomitized the nearby laminated limestone of unit 7 and at the same time were altered dynamically so as to produce a schistose structure, a rock similar to the greenstone of unit 10 might result.

The chlorite schist of unit 3 is composed principally of chlorite and quartz. Minor amounts of calcite, epidote, and orthoclase are found. The chlorite on the basis of its optical properties is near ripidolite in Winchell's chart (1936, p. 643, fig. 1) and, as such, is about midway between the high iron chlorites and the high magnesium chlorites. It could have been derived by the hydration and alteration of almost any ferro-magnesian mineral containing aluminum. As about 50 per cent of free quartz exists, it appears that the original rock contained about 60 per cent silica and therefore could have been a flow rock or tuff of about dioritic composition with small feldspar content. The high iron and magnesium content leads the writers to favor a composition somewhat more basic than diorite.

Crosscutting quartz veins were observed at several places but particularly in the south-central part of the island, where the sample of the chlorite schist of unit 3 was obtained.

ANTELOPE ISLAND

Small remnants of slightly metamorphosed sedimentary rocks are found on Antelope Island resting in sedimentary contact on the truncated edges of the older Farmington Canyon complex rocks (Fig. 2). The

[2] *See* footnote 1.

isolated patch of strata capping the highest peak—units 5 and 6—is believed to overlie the variegated shales or argillites—unit 3—found a short distance down the slope. A large part of the north end of the island is veneered with Lake Bonneville beach deposits, but a few quartzite outcrops are thought to be the Brigham quartzite of Cambrian age. The section is as follows:

Units	Thickness in feet
Cambrian (?) quartzite exposed on Antelope Island.	
7. Quartzite generally pink and white, some purple. Fairly fine and partly sugary crystalline. Several intraformational conglomerates with pebbles of quartz ½ to 3 inches in diameter occur. Badly brecciated in all outcrops. Thickness unknown but probably greater than 500 feet. Probably rests unconformably on Farmington Canyon complex and, before stripped off by erosion, rested unconformably on Promontory phyllites (Fig. 4)...	500+
Island sequence exposed on Antelope Island.	
6. Limestone, gray, laminated. Similar to unit 7 of Promontory Point. With unit 5 forms cap on highest peak of Antelope Island.....	12
5. Quartzite, dark purple. Rests in sedimentary contact on Farmington Canyon complex.....	4
4. Removed by erosion.....	100-200
3. Slaty shale, even laminations 1/16 to 2 inches thick, with colors of purple, brick red, greenish gray, and pinkish brown. Not micaceous. Clay to silty texture. Fairly hard and cleaves like good slate but parallel to bedding. Top removed by erosion (Fig. 2).....	75+
2. Limestone, pinkish gray, finely crystalline. Laminations are visible in places. Shattered in places with network of 1/32 to ⅛ inch quartz veinlets. The unit probably thickens and thins along strike. Marble texture in places.....	31
1. Tillite, light purple gray with green-gray patches. From distance almost black. Matrix of clay to silt texture (now argillite) containing scattered particles of all sizes from sand grains to boulders 5 feet across. No trace of stratification or assortment. Pebbles and boulders consist of red, black, and white quartzite, and gneissic granite. The granite pebbles are the most abundant. Very similar to Blackwelder's Little Mountain tillite. Rests in sedimentary contact on 3-foot weathered layer of Farmington Canyon complex (Fig. 2). (*See* thin-section description that follows.)...	15-30

Under the microscope the matrix of the tillite of unit one—Antelope Island—was found to be composed largely of a fine-grained mixture of quartz and sericite. A considerable percentage of the quartz grains have diameters varying between 0.5 and 0.1 millimeter. Relatively few grains are larger, although some measured 1.5 millimeters in diameter. Minor amounts of other minerals include orthoclase now completely altered, magnetite, ilmenite, rutile, leucoxene, zircon, chlorite, and apatite. The quartz grains have angular to subangular shapes. Metamorphism has developed an imperfect schistosity. The long axes of the quartz grains lie in the plane of elongation of the sericite crystals. This orientation is parallel with and probably the cause of the schistosity. However, no evidence of stratification can be observed. The tillite differs from that occurring at Little Mountain, which was described by Blackwelder (1932,

p. 291-296), by its apparent lack of fresh feldspar, secondary calcite, limonitic material, chlorite, and biotite. Aside from the simpler mineral composition, it fits Blackwelder's description very well.

A thin laminated limestone that caps the highest peak on Antelope Island is believed to correlate with the limestone of unit 7 on Promontory and the dolomite of unit 7 on Fremont Island. The limestone of Promontory Point is dark gray, due to considerable ferrous sulphide. The Antelope Island limestone (unit 6) is tan gray, due to abundant limonite. The chemical state of the iron is their only important difference; they both contain much iron and little magnesium, and their insoluble content is remarkably similar. The insoluble content of the Promontory Point limestone contains 90 per cent of muscovite (Antelope Island limestone, 70 per cent), 5 per cent of orthoclase, microcline, and acid plagioclase (Antelope Island limestone, 25 per cent), and 5 per cent (Antelope Island limestone, 5 per cent) of biotite, chlorite, hematite, magnetite, tourmaline, zircon, rutile, titanite, anatase, leucoxene, and very little quartz. The zircon grains in both rocks are purple and well rounded; the tourmaline grains in both are usually strongly pleochroic and colorless to dark green.

The surface of the Farmington Canyon complex had several hundred feet relief in the Antelope Island area at the time the sediments of the Promontory phyllites were being laid down. The tillite rests on the older schists in one place, and younger beds such as the quartzite of unit 5 rest on the complex in another.

No outcrop was observed in which the contact of the pink quartzite of unit 7 with the Farmington Canyon complex or the Promontory phyllites was exposed. The evidence supporting the correlation of the pink quartzite with the Brigham quartzite is, therefore, indirect and comes from lithologic similarity and structural relationships. The Brigham quartzite has the same color and a comparable thickness but in numerous localities north of Ogden Canyon it is generally slightly coarser than the quartzite of Antelope Island. In the characteristic Brigham quartzite the individual sand grains are fairly distinct megascopically. Some exposures of the Antelope Island quartzite are granular, but others are sugary crystalline and in this respect are similar to some of the pre-Cambrian quartzites in Willard Canyon.

If the quartzite is Cambrian, it probably rests unconformably on all the older rocks. This structural relation fits best the areal geologic map (Fig. 2) and is the strongest evidence of the Cambrian age of unit 7. If the north–south contact of the Farmington Canyon complex and the quartzite is a high-angle fault, it must be broken by hidden cross faults

or curve incredibly in order to miss the Promontory phyllite exposures to the south and the small island called White Rock to the west, which is composed of the marble of unit 2. A study of the Laramide structures of the area also leads the writer to believe that it is improbable that the distribution is due to dissection of a thrust sheet.

LITTLE MOUNTAIN

The rocks of Little Mountain were first described by Stansbury (1853, p. 164):

"A slate is side by side with a dark rock filled with pebbles and stones as large as a man's head . . . appeared to be a granite altered by intense heat . . ."

Blackwelder (1932, p. 291-296) has recently described the rocks on Little Mountain and believes the "dark rock filled with pebbles and stones" is a tillite. His rather detailed description could have been used with little change for the tillite of Antelope Island. It is believed to be the same and makes a good horizon-marker. Blackwelder's section is as follows:

Units	Thickness in feet
2. Tillite. (*See* Blackwelder's description.)	300+
1. Phyllites (called slates by Blackwelder), silvery to dark smoky gray. Evenly laminated to varved. Secondary pyrite cubes in certain zones In upper part of sequence a few beds of massive, hard greywacke interbedded with phyllite. A greenish amygdaloidal lava flow near the top of the unit	600+
Total exposed thickness	900

Blackwelder believes the laminated phyllites are metamorphosed glacial lake clays.

BIRD ISLAND AND OTHER LOCALITIES

Bird Island (Hat Island) is composed of tillite. Blackwelder (1932, p. 299-301) describes this tillite as having the same characteristics as the tillite of Little Mountain, and in this report the two are correlated.

Stansbury (1853, p. 164) described the tillite and phyllite beds on an islet between Little Mountain and Fremont Island, known as "Landing Rocks." Blackwelder believes these rocks are the same as those in Little Mountain.

The "conglomerates" on Stansbury Island mentioned by Stansbury were not seen by the writer, although the sequence of beds was fairly well traversed. Some calcareous shales of slaty appearance occur there which are pre-Madison and Paleozoic. The same shales crop out on the adjacent Carrington Island which Stansbury calls "fine roofing slate", but no pre-Cambrian rocks crop out on Stansbury or Carrington Islands.

CORRELATION

The limestone of unit 7—Promontory Point—the dolomitized limestone of unit 7—Fremont Island—and the limestone of unit 6—Antelope Island—represent the only thin and presumably definite lithologic tie between sequences, but even this correlation is beset with difficulties. The thick tillite is above the dolomite on Fremont Island, and the tillite is below the limestone on Antelope Island. On Promontory Point there is no tillite above or below. However, the tillites, where they occur, are only short stratigraphic distances above or below the limestone, and it is concluded hesitantly that the tillites are lenticular and recurrent in distribution. As a general correlation the upper part of the phyllites on Fremont Island probably corresponds to those on Promontory Point.

If these correlations are approximately correct, then the sediments thicken greatly from Antelope Island to Fremont Island and appear to have been deposited in a deep trough.

NORTHERN WASATCH SEQUENCE

GENERAL CHARACTERISTICS

Thrust faulting of complex nature has rendered stratigraphic studies difficult in the Wasatch Mountains north of Ogden, and as a result no one until now has attempted to compile a complete section between the crystalline complex and the known Cambrian. Blackwelder (1910, p. 517-542) worked out the salient features of the thrust structure, and upon this basis the writer (Eardley, 1939) has done additional field work. In the summers of 1938 and 1939 Lee Christenson, who had been making a thesis study of the northern Wasatch, found that the thickest sequence of beds below the Langston limestone of Middle Cambrian age is exposed in Bakers Canyon, about 2 miles north of Brigham City. By tracing the lowest beds in Bakers Canyon southward along the mountain front, it is evident that they are separated stratigraphically by only a small thickness of beds from the top of a dark argillaceous unit in the mouth of Brigham Canyon. This argillaceous unit may be traced southward into Three-Mile Canyon where it is believed to overlie a quartzite sequence (Fig. 3). At the base of the quartzite sequence is another dark argillaceous unit containing a tillite. The argillaceous unit may be traced southward to the upper north divide of Willard Canyon. From here to the thrust fault near the mouth of the canyon a continuous sequence of quartzites and phyllites 3500 feet thick is exposed. Two thick phyllite units separated by a quartzite make up the base of this sequence. For areal mapping the four argillaceous units and the quartzite sequences between would constitute the most logical lithologic units because they contrast so conspicuously in color and because their exten-

sive outcrops stand out prominently. A total thickness of about 8000 feet of strata below known Cambrian beds is, therefore, present. No angular or disconformable relationship was found between the Brigham quartzite and the underlying beds, and hence the writers cannot point to any division plane between Cambrian and pre-Cambrian strata.

Of greatest interest is the recognition that the Ophir shale and Tintic quartzite formations extend from the southern Wasatch Mountains, and probably their type localities in the Oquirrh and Tintic ranges respectively, northward to Willard Canyon where the Willard thrust has imposed the Proterozoic (?) sequence upon them (cross section, Fig. 3). The Cambrian sequence above the Proterozoic (?) strata—i.e., above the thrust—is different in succession of units, thickness, and fossils. After the various Proterozoic sequences have been described a concluding chapter on the Proterozoic sedimentary setting will be presented in which is concluded that here in the northern Wasatch an eastern, thick, offshore sequence has been thrust westward over a thinner near shore sequence. Instead of an abrupt north–south change in character of Cambrian sediments, an east–west change is recognized in harmony with the conception of highland and trough structure and accumulation of sediments.

BAKERS CANYON

The best exposure of the Brigham quartzite in the Brigham City portion of the Wasatch is in a narrow and steep canyon draining the western mountain front about 2 miles north of the City. The United States Forest Service map designates a small village at this point named Baker. For want of a name for the canyon due east of the village (just a farm house), the name Bakers Canyon is here chosen. The canyon is probably Walcott's (1908, p. 8) type locality for the Brigham quartzite. Northward the Brigham quartzite and Langston limestone contact loses elevation, and less quartzite is exposed. Southward an east–west cross fault disturbs the continuity, and the exposures are poorer. Bakers Canyon was chosen, then, as the best place to measure the strata below the Langston limestone of known Middle Cambrian age. Particular effort was made by Christenson and the senior writer to find and describe a break of some kind between the quartzites designated as Brigham—generally believed to be Middle or Lower Cambrian—and the quartzites and phyllites below, generally considered to be Proterozoic. The section is as follows: (Compare with the graphic columns of Figure 3.)

Units	Thickness in feet
49. Langston limestone. Middle Cambrian	?
48. Quartzite, gray, massively bedded, thin laminae and thin cross-bedded units. Generally coarse-grained with some thin shale partings. An indistinct trilobite 2 inches long was found in a shale parting in the upper 75 feet.	225

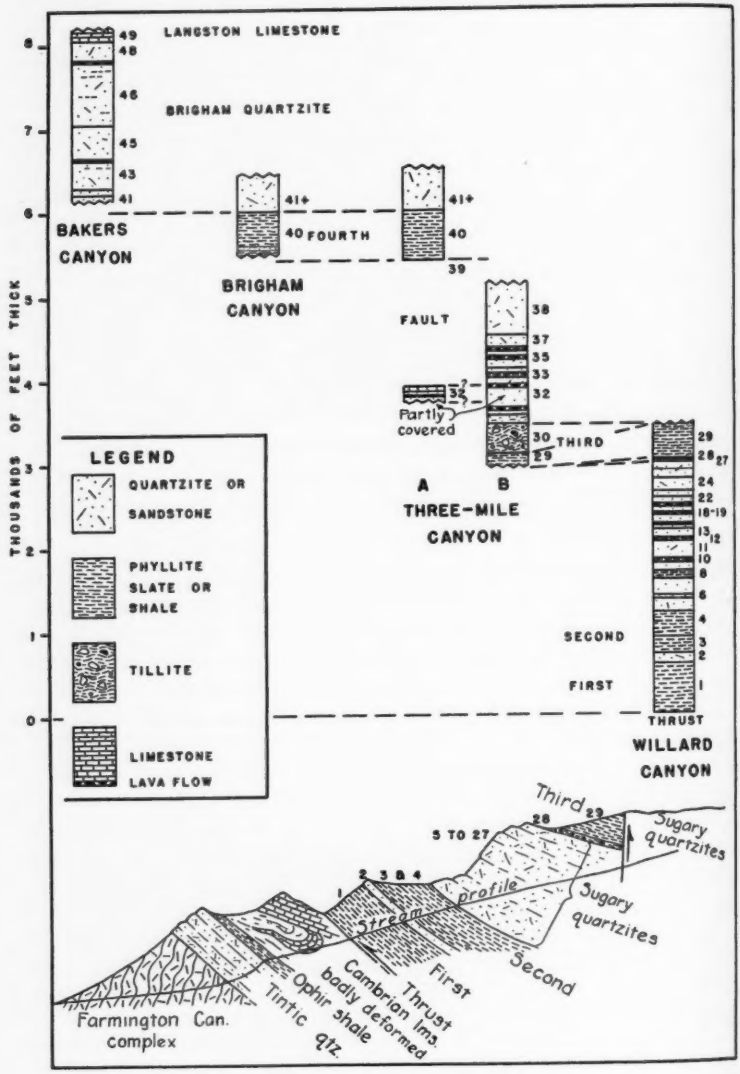

FIGURE 3.—*Northern Wasatch sequence*

Cross section is along the north side of Willard Canyon. The words first, second, third, and fourth refer to the dark argillaceous members within the quartzite sequence. Saw-tooth tops and bottoms of columns indicate that sections continue upward or downward but were not measured.

Units	Thickness in feet
47. Grit, quartzitic. Beds 3 to 6 inches thick, interlayered with green-gray sandy shale. The bedding surfaces are wavy with mica coatings resembling Ophir shale. Some shale beds display thin laminations	20
46. Quartzite, light to dark gray. Some beds almost white. Medium- to coarse-grained with beds 6 inches to 3 feet thick. Micaceous, green-gray shale partings scarce in lower part but numerous toward top of unit	760
45. Quartzite, light pink and almost white at top. Fine- to medium-grained, pure quartz sand. Massive bedding with pronounced cross laminations, either beach or eolian	400
44. Conglomerate and quartzite. Conglomerate beds 3 inches to 3 feet thick interbedded with quartzite layers. Pebbles well rounded and average 1 inch in length. Maximum size 1½ inches. They are composed of 75 per cent vein quartz and 25 per cent quartzite. No discordance above or below conglomerate zone could be detected. Quartzite beds and quartzitic conglomerate matrixlike beds above and below. Concluded that conglomerate beds are intraformational	20
43. Quartzite, generally light pink and massively bedded. Some of pink quartzite beds have gray-green laminations. Green and silvery coatings of sericite or muscovite common on bedding surfaces. Conglomerate layers up to 3 feet thick are common. Pebbles composed of white vein quartz and pink, gray, and purple quartzite. Maximum size 1 inch. Prominent cross-bedding suggests beach deposit because of channeling, cleanness, rounding, and no foresets	240
42. Quartzite, greenish gray and pink, medium-grained and evenly sorted. Beds 1 to 3 feet thick. Some green beds show pink mottling. Conglomerate stringers present. Pebbles well-rounded and composed of white, pink, and gray quartzite and dark red jasper. Maximum size 2 inches. Purple micaceous argillite, 6 inches thick at top of unit	70
41. Quartzite, light pink, medium- to coarse-grained. An intraformational conglomerate 6 inches thick observed. Pebbles average 1 inch in length, composed of white and pink vein quartz and purple quartzite. Base covered with alluvial fan	40+
Total thickness below limestone	1775+

The section is completely exposed, and Christenson and Eardley are sure that no unconformity sufficiently angular to be seen along the outcrops in Bakers Canyon exists. All the conglomerates appear to be intraformational. These conclusions are controversial because Hinds and Marsell visited Brigham Canyon 2 miles to the south, climbed the north face, and found discordances in strike and dip readings which Hinds (1936, p. 84-85) later reported as an angular unconformity between the Brigham quartzite and underlying Algonkian beds. The Brigham Canyon locality was visited by Christenson and Eardley after Christenson had previously carried his studies southward in reconnaissance to this area. They both agreed that, by tracing as well as possible the beds of the Bakers Canyon exposure to Brigham Canyon and also by estimating the thickness of beds below the Langston limestone on a peak to the north of Brigham Canyon, the position of Hinds' unconformity stratigraphically is above the base of the sequence measured in Bakers Canyon. Hinds estimates the "unconformity" to be about 600 feet below the top

of the quartzites. As no unconformity was found in the entire sequence in Bakers Canyon and especially not at 600 feet below the top, Christenson and Eardley searched for the Brigham Canyon discordance. They found shattered quartzite, talus-covered slopes and poor exposures, and a cross fault zone extending from the northeast through a saddle into the north face of Brigham Canyon. It was concluded that any discordance in strike and dip of two sets of quartzite beds in this setting was not sufficient evidence to postulate an unconformity. While recognizing the fact that low-angle unconformities are usually detected by mapping along the contact and are often not observed in cross sectional exposure, it cannot yet be granted that an angular unconformity exists. It should also not be accepted that the stratigraphic position of a break between Cambrian and pre-Cambrian strata, if it exists, has been found in the eastern offshore sequence of the northern Wasatch Mountains. As previously explained, the eastern offshore sequence is that of the overriding block of the Willard thrust, moved, according to the senior writer's (1939a, pl. 1; 1939b, fig. 4) interpretation, from east to west.

BRIGHAM CANYON

The first exposure in Brigham Canyon as it is ascended is composed of dark argillites and phyllites. These may be traced around to the west face of the mountain and northward to a wind gap where a fault drops them on the north side just below the alluvial fans. They are overlain by a series of quartzite beds, which in large part are badly shattered, and are equivalent to the lower part of the sequence in Bakers Canyon. An outcrop of greenish-gray quartzite beds up-canyon from the argillite outcrop is probably equivalent to beds in units 42 or 43 of Bakers Canyon. The sequence in Brigham Canyon was only studied sufficiently to carry the correlations southward to Three-Mile Canyon. It is as follows:

Units		Thickness in feet
41+.	Quartzites, pink and greenish gray, medium- to coarse-grained	500+
40.	Argillites, slates, phyllites, dark-gray and brown to dark purple. Lower wavy phyllites contain purple quartzite beds	500±

THREE-MILE CANYON

Three-Mile Canyon is located on the west front of the Wasatch Mountains 3 miles southeast of Brigham City. Blackwelder (1910, p. 296-297) describes its four major units: an upper thick quartzite series, a dark-gray phyllite, a 300-foot tillite, and an underlying laminated light-gray phyllite.

On the north side of the mouth of the canyon is a high dark tillite cliff. The section (column B, Fig. 3) beginning here and continuing upward

along the crest of the spur to the northeast is believed to be continuous and is as follows:

Units	Thickness in feet
39. Quartzite, purple and gray in lower 50+ feet of unit. Rest of unit not known but estimated to be about	300
38. Quartzite, light gray, massive, cross-bedded, vitreous luster, but grains distinct. Medium- to medium-coarse-grained. Green, white, and flesh-colored varieties interbedded	600
37. Quartzite, buff on fresh surfaces but weathers rusty	150
36. Phyllite, dull gray	25
35. Quartzite, gray, less pure, medium coarse with granular appearance. A thin dark lava flow (?) 50 feet from top	200
34. Arkosite or impure quartzite	40
33. Argillite, dark gray-blue, laminated. A bed probably altered lava flow occurs about 30 feet from top. (*See* microscopic description that follows.)	60
32. Quartzite, light gray and purple beds. Medium-grained. Buff quartzite impure. Some mottled gray beds may be lava flows. (*See* microscopic description that follows.)	450
31. Arkosite with quartzitic and phyllitic beds. Possibly also some greywacke beds or flow rocks	100
30. Tillite. Blackwelder's description follows: Boulders are 60 per cent gneissoid granite, 35 per cent white and gray metaquartzite, and 5 per cent vein quartz and buff dolomite or marble	300-400
29. Slate and phyllite, striped gray and silvery gray. Some gray arkose and greywacke beds. Some black and gray silty phyllites contain pyrite cubes. Base not definite, and relations obscure	200+
Total thickness	2525+

In ascending the canyon about one mile beyond the imposing dark tillite cliff the tillite unit is apparently followed along the north valley wall to a prospect drift which, surprisingly, is not driven in tillite as would be supposed but in various phyllites. The section measured from the mine northeastward up a spur on the north side of the canyon is as follows. The units are described from top to bottom and are shown in column A of Figure 3.

Units	Thickness in feet
41+. Quartzites, pink and light gray. Believed to be equivalent to units 41 to 43 in Bakers Canyon and unit 41+ in Brigham Canyon	500+
40. Siltstone slate, with phyllite variations. Thick dark rusty-weathering unit. Light-gray slates in upper part. Some dark beds in lower part appear carbonaceous and break down into powder	600±
Fault	
32d. Quartzite, granular and coarse, rusty stringers of cement. Some finer beds and somewhat sugary crystalline	15
32c. Limestone, pink and tan beds, partly laminated. (*See* following description.)	6
32b. Sandstone calcareous tan and gray, beds ⅛ to 3 inches thick. Fine- to medium-grained. Quartzite, phyllite, and laminated shale variations	30
32a. Limestone, argillaceous, laminated. Greenish-gray dominant color, but some beds flesh-colored. Thin beds of fine quartzitic sandstone interlayered. (*See* following chemical description.)	50+
31. 30, 29? Phyllites in prospect drift. Started in rusty speckled arkosic phyllite and ended 200 feet in, in black carbonaceous and graphitic phyllites. Some phyllite fragments on the dump are finely laminated, appear soapy, and are composed chiefly of sericite	200+

The fault is believed to bring the argillaceous unit of Brigham Canyon down into contact with beds immediately overlying the tillite and associated beds of Three-Mile Canyon. Inasmuch as the fault runs N 30° W, the limestone units below the fault are apparently continuous to the west with unit 32 on the spur above the tillite outcrop. It was concluded in the field that the limestone beds were covered and missed in measuring the section up the crest of the tillite spur and probably belong somewhere within the 450-foot quartzite unit.

The rocks in Three-Mile Canyon, as implied from the above discussion, are partly concealed and further complicated by faulting, but the sequence arrived at by correlating the two sections described above essentially agrees with the major units mentioned by Blackwelder and adds considerable detail. Although the above sequence may be altered when detailed mapping is carried out, the changes will probably not be great, and for purposes of approximate correlation of the thick pre-Cambrian sequence of strata in north-central Utah the compiled section is sufficiently correct.

The limestone beds are the best markers, presumably, for correlation purposes, because of their unusualness, their similarity, and their association with tillites. A specimen from 32c gives a strong test for iron, part of which is in the ferrous state, and shows a small amount of magnesium. The insoluble residue is abundant, clean, and fine-grained. Muscovite constitutes about 80 per cent of the residue, and quartz 20 per cent. A few grains of leucoxene, tourmaline, chlorite, rutile, hematite, and zircon were seen. This rock is very similar in all respects to the limestones of Promontory Point and Antelope Island.

A sample of limestone from unit 32a shows a fair amount of iron, very little of which is in the ferrous state, and a faint trace of magnesium. The insoluble residue comprises 85 per cent muscovite, 10 per cent green chlorite, and 5 per cent of leucoxene, quartz, orthoclase, plagioclase, tourmaline, rutile, zircon, and magnetite. It is not similar in color, but fairly alike in texture and composition to the other limestones just mentioned.

Under the microscope the rocks listed as lava flows in units 35, 33, and 32 are found to have the following similar characteristics: (1) Extensive replacement of quartz, hornblende, and possibly augite by calcite-units 15 and 17 about 40 per cent replaced, and unit 18 about 75 per cent; (2) evidence of a porphyritic texture; (3) phenocrysts, of biotite or its alteration product, chlorite, 0.5-1.0 millimeter in diameter; (4) groundmass, moderately fine-grained, consisting essentially of quartz and some biotite; (5) long needles and short prisms of apatite, 0.25-1.0 millimeters in length, occurring as inclusions in biotite and in rhombs of calcite which appear to be replacements of hornblende and in places possibly augite;

(6) limonitic inclusions and stains are abundant; (7) magnetite, primary and secondary, the latter due to alteration of ferromagnesian minerals, is common; (8) rutile and leucoxene in small granular aggregates are abundant in all rocks except those of unit 32.

The somewhat porphyritic habit and the development of apatite crystals strongly suggest an igneous origin for the rocks. Calcium bicarbonate solutions were probably responsible for the extensive leaching of alkalies and silica and the replacement of some quartz and ferromagnesian minerals by calcite. Only traces of feldspar remain. A secondary chlorite is abundant in the groundmass. The alumina and silica of the chlorite were possibly derived from the feldspar, and the iron oxide and magnesia from decomposed ferromagnesian minerals.

The rock of unit 33 is unusual in that it contains what appear to be grains of quartzite evenly distributed through the groundmass. These grains (?) are about 0.25-1.0 millimeters in diameter and are usually oval, although some are round and others rectangular. An apparent reaction rim of biotite and granular magnetite borders every grain (?). Perhaps the grains (?) were some other mineral that reacted with the magma, was later replaced by quartz, and still later in places by calcite.

The lack of feldspar, which is probably due to leaching and replacement, makes it difficult to classify these rocks. The abundance of biotite phenocrysts and biotite and quartz in the groundmass, however, indicates that they are igneous and of a siliceous character between rhyolite porphyry and dacite porphyry in composition. In view of their extensive alteration the writers prefer to regard these igneous (?) rocks as flows of Proterozoic age and not sills of some later age.

WILLARD CANYON

The rocks traversed in ascending order from the mouth of Willard Canyon are: Farmington Canyon complex, Tintic quartzite, Ophir shale, Cambrian limestone, and the phyllites and quartzites of the Proterozoic (?). The relations are shown in the cross section of Figure 3. The units of the cross section are numbered the same as their equivalents in the columns. The phyllite units 1, 3, and 4 are best measured along the bottom of the canyon, but the quartzite sequence above—units 5 to 15—is freshly exposed in a near-vertical scar of a landslide on the north canyon wall. From the upper rim of the landslide cliff the section was measured along a ridge to the end of a CCC road. This ridge divides the drainage into Three-Mile Canyon on the north and Brigham Canyon on the south. The phyllite and thin tillite of units 29 and 28, as seen from the triangulation signal on the top of the high ridge, form a fairly con-

tinuous dip slope to the tillite and phyllite units in the bottom of Three-Mile Canyon and are therefore correlated with them. From top to bottom the section is as follows:

Units	Thickness in feet
29. Phyllites, dark gray, and in part laminated. Arkosic and purple quartzitic variations. Thin layer of tillite near top	400
28. Tillite. No large boulders observed	20±
27. Quartzite, gray-white	90
26. Quartzite, purple	20
25. Quartzite, gray-white	90
24. Quartzite, gray to white	150
23. Schist, muscovite and quartz muscovite varieties. Cut by quartz veinlets	65
22. Quartzite, gray and white. Hematite vein crosscutting, 6 inches thick	60
21. Phyllite, dark and massive, thin white bedding lines	20
20. Quartzite, gray and white with thin chlorite schist beds	50
19. Conglomerate (?). Angular fragments of dark chlorite magnetite schist, some with limonite spots probably once magnetite. Foliation of pebbles generally parallel, but some foliated pebbles conspicuously not in plane. Quartz matrix is like vein quartz but displays sheeting. Some fragments have replacement contacts as if matrix is a secondary replacement. Some large gray quartzite boulders (?) observed	10
18. Grit, light gray. The quartz grains appear euhedral	2
17. Metaquartzite,[a] grayish white, vitreous, and sugary crystalline. Numerous quartz veins 1 to 2 inches wide	80
16. Phyllite, mottled gray green with scattered quartz grains. Epidote veinlets	25
15. Quartzite, white and buff	35
14. Quartzite, hard speckled, dark steel gray with thin interbedded phyllites, rich in hematite	5
13. Metaquartzite, white, fine sugary crystalline. Weathers rusty. Pure specular hematite in veins filling joints, 2 inches thick near base. Two gray phyllite beds 2 feet thick near top of unit	85
12. Chlorite schist with thin lenses of quartz, dark, thick-bedded with massive gray quartzite beds. Foliation at angle to bedding	35
11. Metaquartzite, white, sugary crystalline	100
10. Metaquartzite, pale jade green and translucent. Fine-grained and sugary crystalline on certain fracture surfaces. Slightly schistose on others. (Microscopic description follows.)	20
9. Metaquartzite, same as unit 11	100
8. Phyllite, dark gray. Not seen closely	30
7. Metaquartzite, white, finely crystalline. Weathers rusty. (Microscopic description follows.)	200
6. Metaquartzite, purple	15
5. Metaquartzite, pink and white, sugary crystalline	150
4. Phyllites, silvery, laminated. Also considerable dark phyllite, possibly carbonaceous. Some phyllites border on sericite schist	300
3. Phyllite, gray	200
2. Metaquartzite, white, sugary crystalline. Weathers rusty pink	100
1. Phyllite, hard, rusty-purple appearance at distance. Base cut off by Willard thrust. Detailed sequence of subunits from top to bottom is as follows:	
j. Phyllite, tan gray, sandy. Beds 1 inch thick.	
i. Phyllite, mottled gray and dark gray. Some layers sandy.	
h. Quartzite (?), rusty, coarse, impure.	
g. Phyllite, green, chloritic, smooth.	
f. Phyllite, purple gray.	
e. Slate, silvery gray surfaces with phyllitic tendencies.	

[a] *See* footnote 1 for definition of term.

Units	Thickness in feet
d. Phyllite, green, chloritic. Scattered black specks of biotite.	
c. Phyllite, gray, granular texture.	
b. Phyllite, purple gray.	
a. Phyllite, gray, considerably sheared.	
Total thickness of unit 1	600
Total thickness	3057±

The meta-quartzites of units 2, 5, 7, 10, 11, 13, 17, and others below 28 are called such because their grains have been shattered and partially rehealed, with strongly sutured intergrown boundaries. Slightly elongated grains with parallel axes contribute a faint megascopic schistosity to the rock. At right angles to the schistosity, however, fracture surfaces present a pure sugary appearance. A small amount of green chlorite in disseminated shreds imparts a pale-green color to the rock of unit 10.

Unit one is well exposed and most accessible on the high ridge southwest of and at the end of the high road from the CCC camp. The above detailed description is from this locality.

The first and second phyllites at the base of the section (cross section, Fig. 3) were traced to the head of Willard Canyon just above and southward of the CCC camp, and no beds stratigraphically lower than those in the bottom of the canyon appeared along the thrust surface. It cannot be told, however, how far the crystalline complex was below these units at the time they were deposited or before they were thrust westward over the very thin near-shore sediments onto the highland margin.

OGDEN CANYON

At the head of Ogden Canyon, on either side of the canyon neck at the new Pineview Dam, are road-cut exposures of phyllitic and arkosic rocks. They are badly broken, and it is difficult to make out the bedding. Fresh boulders of silvery phyllite, together with arkose, and greywacke boulders are numerous in the hollows draining the east slope of the Wasatch from Ogden Canyon northward to Liberty. The rocks have all the characteristics of the Promontory Point and Willard Canyon phyllites, but just where they fit into the sequence is unknown.

The arkose is yellowish brown and is composed chiefly of angular to subangular grains of quartz about 1.4 to 0.2 millimeters and less in size. Orthoclase and plagioclase feldspar grains are about the same size as the quartz but vary considerably in amount in different specimens. The average grain size is about 0.3 millimeters. The cementing agent between the grains is a crystallized mixture of sericite, extremely fine quartz, a green biotite, and chlorite. In the specimens studied the quartz grains

are strained, and in one their long axes lie in parallel planes with the sericitic and biotitic material elongated parallel to the same planes. This structure may be the result of slight metamorphism or it may represent bedding plane laminations. In addition to the minerals already mentioned, small amounts of magnetite, limonitic material, apatite, zircon, rutile, and leucoxene are distributed through the rocks.

OTHER LOCALITIES

A thick section of quartzite beds is exposed in the Middle Fork of Ogden River, but time was available for only a cursory examination by the writer. On the basis of color the sequence is divisible into three units.

Units	Thickness in feet
3. Quartzitic sandstones generally purple. Some thin interbedded shales....	3000
2. Quartzitic sandstones, light pink and gray. Thin group of dark quartzite beds in middle.....	700
1. Quartzitic sandstones, rusty weathering. Many thin beds of dark maroon shale, some of which show sun cracks and ripple marks and are micaceous on bedding planes. Sandstones contain many thin layers of fine conglomerate. In a few beds the pebbles range up to 3 inches in diameter and consist of quartzite. (Description from Blackwelder's notes.).....	3500+
Total thickness exposed.....	7200

Unit 3 forms the entire mountain between Wolf and Geertsen creeks. The beds appear somewhat similar to the Uinta quartzites.

A shale and quartzite, presumably Cambrian, are exposed in the canyon of the South Fork of Ogden River and dip in the same direction as the quartzite beds of the Middle Fork, but whether the two sequences are conformable was not determined.

CORRELATION

The basal unit from the canyon on the north was traced, usually with assurance, to the upper part of the section in the canyon to the south, and upon this basis of lithologic continuity local correlations were made, and a composite sequence was built up for the northern Wasatch. The limestone beds of unit 32 afford the best tie westward to the Island sequence. As in the northern Wasatch, the limestones on Antelope Island are directly above the tillite. The order is reversed, however, on Fremont Island, but still the two unusual lithologic units are stratigraphically close together, and an approximate correlation can be made. (*See* Figure 5.)

The probable lava flows of Three-Mile Canyon substantiate the correlations on the basis of the limestones. In both the Promontory Point and northern Wasatch sequences they occur above the limestone beds.

The upper limit of the phyllites—unit 11—of Promontory Point may correlate approximately with the upper limit of the main argillaceous sequence—unit 36—in the northern Wasatch Mountains. (*See* Figure 5.)

COTTONWOOD UPLIFT SEQUENCE

GENERAL CHARACTERISTICS

Walcott's (1891, p. 319) and Hintze's (1913, p. 91-103) description of the quartzites in the Cottonwood uplift southeast of Salt Lake City is briefly summarized as follows:

The total thickness is about 11,000 feet. The most abundant rocks are quartzites (74 per cent); the remainder includes phyllite, slate, and argillite (26 per cent). The phyllites and slates occur in thick units in the quartzite series and as parting members between massive quartzite beds. The quartzites are mostly light shades of pink, yellow, and light brown; some beds are white. Toward the top of the section thin beds of conglomerate occur whose pebbles are rounded and composed of quartz and quartzite. The quartzite strata are generally prominently cross-bedded with the cross-bedded units small and regularly spaced between true parallel bedding planes. Ripple marks are common. According to the definition given by Tyrrell (1926, p. 305), the argillaceous rocks are chiefly phyllites, or argillites with a phyllitic appearance, and hand specimens of the beds were thus classed by Frank Grout, W. F. Hunt, and T. S. Lovering. The phyllites, slates, and argillites are olive green and dark purple to black. Some of the beds in the lower part of the series are mud-cracked.

A tillite (Hintze, 1913, p. 91-103, Blackwelder, 1932, p. 297), 600 feet from the top of the Big Cottonwood quartzites, has been reported by Blackwelder as similar to the Little Mountain tillite.

The type section has been considered as Big Cottonwood Canyon since the time of the Fortieth Parallel Survey. This canyon presents the greatest sequence in the uplift, but critical information about the basal Cambrian unconformity and the upper part of the sequence comes from Little Cottonwood, American Fork, and Neffs canyons.

BIG COTTONWOOD CANYON

The lower part of the sequence in Big Cottonwood Canyon as described by Walcott (1891, p. 319) was found somewhat in error and is emended in the following list of units. (*See* Figure 5.) The upper units—22, 23, and 24—which include the tillite and quartzitic sandstones below the unconformity and unit 25 above the unconformity, were recognized by Blackwelder, Hintze, and Schneider. These units supplant Walcott's units 13 and 14. The unit numbers in parentheses in the column are Wal-

cott's and were left as he described them. The sequence from top to bottom is as follows:

Units	Thickness in feet
Tintic quartzite	
25. Quartzite, white, well-bedded, containing sheets of conglomerate. A basal conglomerate composed of well-rounded quartzite and gneiss boulders and an unconformity separate the quartzite from the Big Cottonwood quartzites. The quartzite is overlain by the Ophir shale of Middle Cambrian age.	700
Unconformity	
24. Slate. Mentioned by Hinds (1936, p. 82)	100 ?
23. Sandstone, quartzitic, purple	600
22. Tillite. (Described by Blackwelder, 1932, p. 289-304.)	10-100
21. (12) Sandstone, quartzitic, light gray. A number of thin intraformational conglomerates. Pebbles well-rounded and composed of quartz and quartzite.	1500
20. (11) Sandstone, arkosic and quartzitic, purplish and reddish brown	75
19. (10) Sandstone, quartzitic, light gray	700
18. (9) Shale, black, quartzitic, slightly micaceous	75
17. (8) Quartzite, generally gray with buff and olive varieties at base. Beds range in thickness from 2 inches to 10 feet. Some beds have visible sand grains, but others are glassy without distinct granular texture	2700±
16. Quartzite, light tan, thick-bedded	50
15. Slate, dark brown	175
14. Quartzite, rusty	100
13. Slate, brown, olive green, purple red, dark gray and black varieties. Most bedding surfaces are wavy, but black variety is fairly good slate. Phyllitic tendencies conspicuous. Thin white quartzose laminations in some beds. One gray bed with black blotches probably chlorite	40
12. Quartzite, rusty. Unit appears to lens out	35
11. Slate, dark purple brown. Fresh samples are dark gray and laminated 1/16 to 1/4 inch thick. In part fairly good slate	600
10. Quartzite, rusty	25
9. Slate, light brown	60
8. Quartzite, rusty pink on weathered surface. Mostly white, green gray, and pink on fresh surfaces. Beds generally massive. Medium-grained with grains indistinct but without fused appearance	600
7. Slate, brown	15
6. Quartzite, same as unit 8	800
5. Slate, purple	50
4. Quartzite, buff. Weathers rusty	75
3. Siltstones and argillites, conspicuous dark-purple and light-green varieties with purple predominating. Beds fairly thick but break down into fissile slopes. Some light-green beds laminated with thin quartzose layers. One bed is entirely clay and chlorite and would generally be called shale	600
2. Quartzite, generally gray, but some beds greenish, and some light pink. Weathers rusty pink. Fine- to coarse-grained and indistinctly granular	1000±
1. Phyllites, dark purplish grays predominate. Mostly thin-bedded. Network of ridges on some surfaces suggests deformed mud cracks. Some thin beds of dark purplish gray chloritic quartzite	1000+
Total thickness below unconformity	11075+

ROCK CANYON

In Rock Canyon, northeast of Provo, the Tintic quartzite rests probably unconformably upon a tillite, at least 300 feet thick, whose base is not exposed (Schneider, 1925, p. 38-40; Blackwelder, 1932, p. 298-299; Hinds,

1936, p. 82). The top of the tillite is slightly weathered and, according to Blackwelder, is similar to the Little Mountain tillite. He lists the percentages of various kinds of boulders in the tillite as follows:

	Per cent
Dense gray dolomites, some finely laminated	45
Metaquartzites, white, brown, and dull gray	50
Miscellaneous rocks, including chloritic and sericitic phyllites, hornblende schist, and gneissic granitoid rocks	5

All the boulders except those of dolomite could have come from the Farmington Canyon complex. The dolomite on Fremont Island is similar, but time and place relations are obscure.

The similar position of the tillite of Rock Canyon and the tillite in the Cottonwood uplift just below the Cambrian unconformity leads the writer to believe that they are parts of the same formation.

UINTA MOUNTAINS (WEST END)

The quartzitic and argillaceous beds forming the core of the Uinta Mountains have been mentioned many times in the literature, but no one has made a detailed section of them. They have been correlated with the Cottonwood uplift sequence by most writers. The present writers studied partial sequences of the strata in the upper reaches of the Provo River from Soapstone Camp to Mirror Lake on a trip in 1939 and report the following inadequate section. The thicknesses were determined graphically from strike and dip readings and map distances and elevations.

Units	Thickness in feet
4. Quartzitic sandstone below Lodore shale, supposedly Lower Cambrian	1000
3. Quartzitic sandstone, purple, white, and pink, coarse-grained. Probably other varieties not observed	4800
2. Shale, in part black fissile, in part micaceous, with Ophir shale appearance. A few thin quartzitic sandstones interbedded	200±
1. Quartzitic sandstone, light gray, pink, speckled white, tan, and thinly laminated green varieties. All generally cross-bedded. Little greenish-gray micaceous shale in Mirror Lake section. Only part of unit observed	6000
Total exposed thickness below Lodore shale	12,000±

According to Powell (1876, p. 144) the upper quartzite unit in the eastern part of the range is separated from the lower units by an unconformity. According to Lovering this observation has been confirmed by later United States Geological Survey geologists. Hintze (1934, p. 166) reports an unconformity at the base of an upper quartzite unit 1000 feet thick in Smith and Morehouse Creek in the west end of the range, but Schneider, Forrester, and the writer visited the same locality together and could not be sure of the existence of such a break. If an unconformity occurs in the upper part of the thick quartzite sequence in the western end, it must be of very low angle.

General observations by Forrester (1937, p. 638) on the lithology of the sequence are as follows:

"Numerous beds of black . . . shale are intercalated in the series. . . . Gradations from true quartzite to sandstone may be seen throughout the range, and in many places the quartzite beds directly overlie the sandstone layers. . . . Near the top . . . the beds become arkosic."

CORRELATION

In the field the strata of the west end of the Uinta Mountain below known Cambrian beds appear less quartzitic and less phyllitic as a whole and contain a smaller per cent of argillaceous beds than those in the Cottonwood uplift. Hand specimens of the quartzites in the laboratory are very similar. Their stratigraphic position below the Middle Cambrian and above a crystalline complex relates them as a whole, but within the sequences no correlation ties were discovered. The shale unit 2 in the Uinta sequence may possibly correlate with the upper part of the argillaceous series in Big Cottonwood. This is hesitatingly suggested in the correlation chart (Fig. 5).

The tillite near the top of the section in Big Cottonwood Canyon probably correlates with the tillite of Rock Canyon. Although these tillites are similar lithologically to those of the northern Wasatch and to Antelope and Fremont islands, it does not seem possible that they were deposited at the same time because of their widely separated positions in the two stratigraphic columns. It is possible that they correlate with the thick tillite in the Sheeprock Range.

SOUTHERN WASATCH SEQUENCE

DRY MOUNTAIN

Dry Mountain is east of Santaquin in the southern Wasatch Mountains. Hintze (1913, p. 97) found a conspicuous angular unconformity there between the Tintic quartzite and underlying quartzites and shales. The unconformity is characterized by a basal conglomerate in the Tintic quartzite and an angular divergence of 15 to 25 degrees of the overlying and underlying beds. The beds under the unconformity thicken from 500 to 1000 feet in 3 miles because of truncation at the old erosion surface. The section was briefly described by Loughlin (1920, p. 324) and later by Eardley (1932, p. 311-314).

The rocks below the Tintic quartzite are interbedded quartzites and shales. The quartzites are chiefly dark red to purple and range from fine-grained to coarse. Most beds are firmly cemented and very hard, but some less well-cemented sandstones occur. Intraformational conglomerates of white quartz pebbles set in a dark-red matrix are common. The conglomerates fracture cleanly through the pebbles. The shales are

chiefly dark red also, but some green and light-yellow beds occur. Some of the shales weather into small angular fragments that form soft talus slopes; others exhibit markings like worm borings.

The quartzites and shales rest in sedimentary contact on the Farmington Canyon complex. A basal conglomerate 15 feet thick is composed of angular and rounded quartz pebbles embedded in a matrix of coarse quartz sand, predominantly light green but in places dark red.

CORRELATION

No distinctive lithologic units were noted within the sequence to tie with other sections. The abundance of shale suggests, perhaps, a correlation with the lower, more argillaceous part of the sequence in Big Cottonwood Canyon. The unconformity would also correlate with similar unconformities in other sections.

SHEEPROCK UPLIFT SEQUENCE

LOCATION AND GENERAL CHARACTERISTICS

The Sheeprock uplift centers about 60 miles south–southwest of Salt Lake City and includes parts or all of the Sheeprock, Simpson, and West Tintic ranges. Its limits and structural details have not yet been worked out, but the general setting in relation to the regional structures has been described (Eardley, 1939a, pl. 1; p. 1282-1284).

The core of the uplift is composed of stratified rocks in large part similar to the Proterozoic (?) rocks of the Wasatch. In the examination of small mining districts Loughlin (1920, p. 424-425, 430) observed a thick sequence of "dark to light brownish quartzite, shale, and conglomerate of probable pre-Cambrian age." The quartzites are commonly chloritic, and the shales

> "have lost much of their shaly character, and the terms hornfels and slate are appropriate to some of them . . . or even phyllite. . . . The conglomerate is typically unassorted and is composed of subangular pebbles of quartzite, metamorphosed shaly rock, granite, and gneiss in a matrix of shaly character. The pebbles or cobbles are prevailingly large, being 4 to 12 inches or more in diameter. The largest noted was an elliptical boulder of quartzite 4½ feet long. The irregular size and distribution of the pebbles and the shaly character of the matrix suggest a glacial origin of the conglomerate."

The conglomerate that Loughlin recognizes as of glacial origin is in all respects, except for composition of pebbles and boulders, the same as the tillites of the central and northern Wasatch. The main tillite unit, in the Sheeprock Range, however, is the thickest seen by the writer anywhere in Utah. As measured south of Bennion Peak it is 1100 feet thick.

SHEEPROCK RANGE

The Sheeprock Range of earlier maps is now more generally referred to as the Vernon Division of the Wasatch National Forest (Fig. 4). The

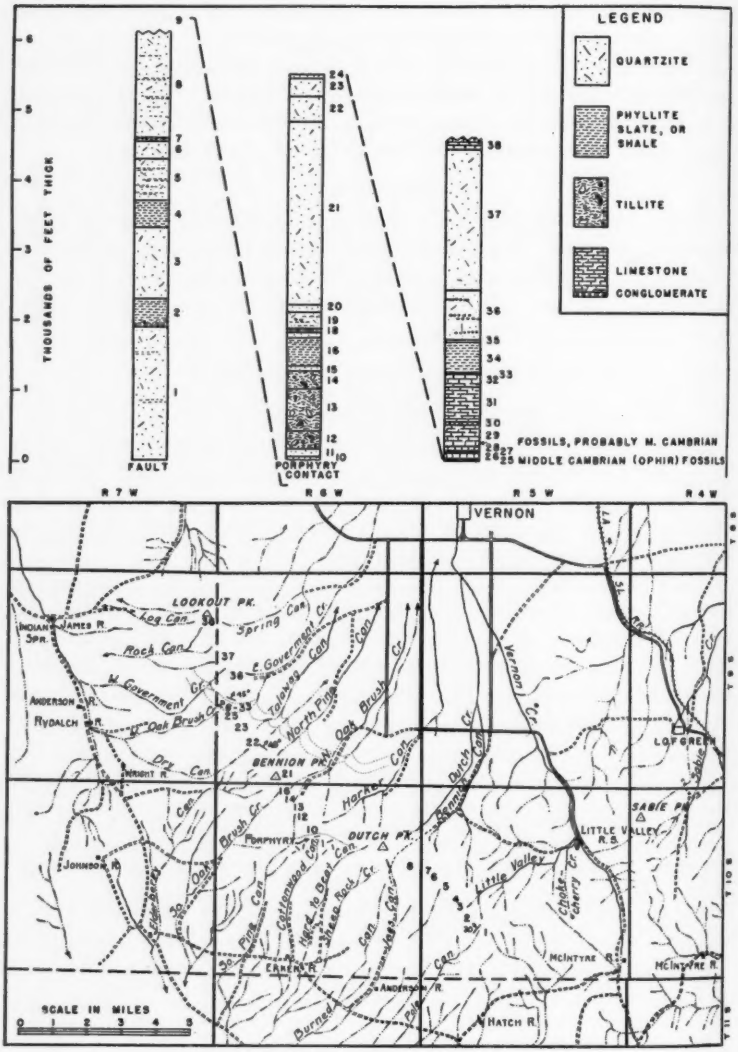

FIGURE 4.—*Sheeprock uplift sequence*

Columnar section and map showing location of described units.

writer began to measure the section at the head of Little Valley southwest of Little Valley Ranger Station. The beds labeled unit number one are undoubtedly not at the base of the section nor are they the lowest beds exposed. Outcrops southward that appeared lower stratigraphically were less favorably exposed for measuring, and the structure seemed more complex. A beginning was, therefore, made at the point where the greatest amount of reliable information could be gained in a limited time. The sequence was measured upward stratigraphically along the range crest northward by pacing traverses, which were corrected graphically for various strike and dip readings.

In the vicinity of Dutch Peak the structure could not be worked out satisfactorily in the time available, and a continuous sequence from unit 8 to 10 was not determined with certainty. A granitic stock also complicates the setting here. However, unit 10 is definitely above unit 8, and it is believed that only a few hundred feet of strata (not exceeding 1000 feet) is unaccounted for. This hiatus is labeled unit 9. The section from top to bottom is as follows:

Units	Thickness in feet
38. Limestone, dark gray. Only seen from distance	?
37. Quartzite, gray, pink, purple, generally coarse and gritty. Some beds not very well cemented, some very hard. Shale at top	2000±
36. Quartzite, purple, medium- to coarse-grained. Intraformational conglomerates common. Upper 350 feet less well cemented and generally gritty. Conglomerate pebbles well-rounded and ½ to 1 inch long, mostly vein quartz but few gray quartzite pebbles. The pass at head of East Government Creek marks top of unit 36	700
35. Shale, very purple, micaceous surfaces	50
34. Shale to slate, olive tan. Beds range from fine clay slate to sandy shale. Some olive-gray quartzite beds intercalated	425
33. Limestone, light gray. Breaks down into small pieces	25
32. Limestone, dark gray, irregularly blotchy	175
31. Limestone, gray, in part crystalline, generally massively bedded, weathers sandy	500
30. Limestone, blotchy gray, weathers creamy tan	30
29. Limestone, black to dark gray, massively bedded. Oölitic beds in upper part. Fragmentary trilobites, possibly *Elrathia* or *Ehmania* and *Glossopleura*	300
28. Limestone, dark gray, thin tan layers alternating. Also much olive-drab-gray shale	60
27. Limestone, gray, massive	15
26. Limestone, black. Beds 1 to 8 feet thick. Irregular tan stringers inclosing ¼- to ½-inch spheroidal bodies	80
25. Shale, olive tan, smooth, nonmicaceous. Trilobites and brachiopods as follows: *Glossopleura producta* (Hall and Whitfield), a species of *Elrathia*, and a species of *Lingulella*, probably *L. ella* (Hall and Whitfield)	25
24. Shale, olive tan, hard, sandy, micaceous. Thin drab quartzite beds intercalated	75
23. Quartzite, light gray, fine to medium. Weathers very rusty	250
22. Quartzite, light pink white is dominant; also pink and olive gray. Generally fine-grained, with sugary appearance	350
21. Quartzite, light purple, medium- to coarse-grained, beds 1 to 4 inches thick	2600±
20. Quartzite, pink, coarse- to fine-grained, uneven texture	100

Units	Thickness in feet
19. Slate and quartzite, interbedded olive-tan slate and creamy olive-tan quartzite. Small flutings and mottles on bedding surface	225
18. Conglomerate. Large boulders in a siltstone slate. Drab to olive tan. Possibly a tillite. No angular unconformity observed	50±
17. Quartzite, light pink, fine-grained, granular but slightly glassy	75
16. Phyllite, dark gray changing to light gray near top. Thin bedded and fissile.	400
15. Quartzite, light gray, fine-grained	75
14. Tillite, brown to dark-gray matrix—looks purple in distance. Matrix has quartzitic tendencies. (Composition of boulders will be described later.)	250
13. Tillite, light-gray quartzitic matrix. Pebbles and boulders are mostly broken homogeneously with matrix to produce smooth joint surfaces. Lenses of quartzite, quartzitic grit, and quartzitic conglomerate occur	600
12. Tillite, purple matrix. Weathers to saddle	250
11. Quartzite, in part friable sandstone, clean whitish tan	100
10. Quartzite, ocher to brownish gray. Weathers dark brown. Generally laminated. Near contact of Laramide stock is micaceous and in part schistose.	50
9. Hiatus. Complex folding and Laramide granite intrusion. Succession not worked out	500±
8. Quartzite with some interbedded phyllites. Gray beds dominant, some brown.	1500±
7. Conglomerate with intercalated quartzite layers. Conglomerate in part stratified and hence water-laid, but in part so irregular as to resemble fanglomerate or perhaps tillite. Has sandy matrix with boulders less than 18 inches long. Foliation of boulders not in common plane. Pebbles and boulders are granite, gneiss, quartzite, and phyllite	50
6. Quartzite, light gray, medium-grained	250
5. Phyllites and quartzites interbedded, dark purple	600
4. Phyllite, gray, partly speckled white and varved. Weathers rusty brown. Cleavage nearly at right angles to bedding	400
3. Quartzite, light gray, generally massive. Medium-grained with granular texture. A green variety of quartzite occurs somewhat like unit 10 of Willard Canyon. A porphyry sill in quartzites of this unit occupies a saddle here. Porphyry similar to Laramide stock of unit 9	1000
2. Phyllite and shale, gray and dark gray. Some conspicuously varved specimens are silty argillites. A thin tillite occurs at the base, and several thin quartzite beds are intercalated	400
1. Quartzite, dominantly gray, fine-grained and granular with glassy luster	1900±
Measured thickness below Middle Cambrian fossils	12,000±
Measured thickness including quartzites above Middle Cambrian fossils	16,500±

The fossils of units 25 and 29 were identified by Resser (personal communication) as Middle Cambrian and identical with those in shale beds in the Oquirrh Range and in Big Cottonwood Canyon. They are also equivalent to the Chisholm to the south and west.

The tillite is divided into three units whose identities are not altogether clear, but, on the basis of the quartzitic matrix and lighter color of the middle part in contrast to the argillaceous and dark purple of the upper and lower parts, a distinction has been made. No marked difference in boulder composition of the three parts was noted, and therefore it must remain speculative as to whether three ice advances are indicated.

The boulders and pebbles identified are as follows: white and pale-green metaquartzite or pegmatitic quartz, granite, gray-buff limestone, and sericite-chlorite schist. The smaller but visible grains are chiefly quartz and microcline.

The matrix under the microscope consists essentially of angular to subangular quartz grains, ranging in size from 1 millimeter downward and imbedded in a fine-grained groundmass consisting of sericitic, chloritic, and probably clay material, as well as quartz flour. Most of the quartz is strained, and some grains were definitely derived from a quartzite. A small amount of microcline and orthoclase also occurs as grains imbedded in the groundmass. Grains of material now entirely altered to sericite or chlorite are common. The groundmass has a distinctly schistose structure, the long diameters of the larger grains also lying in the plane of schistosity. Some of the schistose layers of the groundmass are brown, due to ferruginous material. The matrix, therefore, is fairly similar to that of the other tillites farther north. The cross-bedded sand and gravel lenses in the tillite are particularly good evidence of morainal deposits with their associated fluvioglacial material. Another observation that substantiates the glacial origin of the deposits is the nonsedimentary nature of the pebbles and boulders in the tillite. They came in large part from land areas comprising the ancient Farmington Canyon complex rocks. It is difficult to see how some process such as submarine landsliding could have transported boulders from land or shore areas to sites of clay and sand accumulation, but on the other hand it is simple to understand glaciation as such a process.

CORRELATION

The fossils of Ophir formation affinity, of course, provide an accurate tie to other sections previously described and to the House Range section (Deiss, 1938, p. 1124-1141). Precise correlations of units below the fossil zone, however, are impossible.

The peculiar conglomerate of unit 18 is the only suggestion that the writer saw of a break in the sequence below the Middle Cambrian. In the exposures along the crest of the range no angularity between beds above and below the conglomerate was observed. However, this is not sufficient to disprove the presence of an angular unconformity at the base of the conglomerate, but until one is found it probably is safest not to postulate its existence in any theory of age or origin.

The thick tillite is approximately in the same stratigraphic position as the tillites of the northern Wasatch and the Island sequences and is, therefore, believed to be of approximately the same age. No limestone

beds either above or below the tillite were found, however, to tie with those of the northern sequences.

It is perhaps noteworthy that four great sequences have about the same total exposed thickness below the Middle Cambrian (Fig. 5). Only one

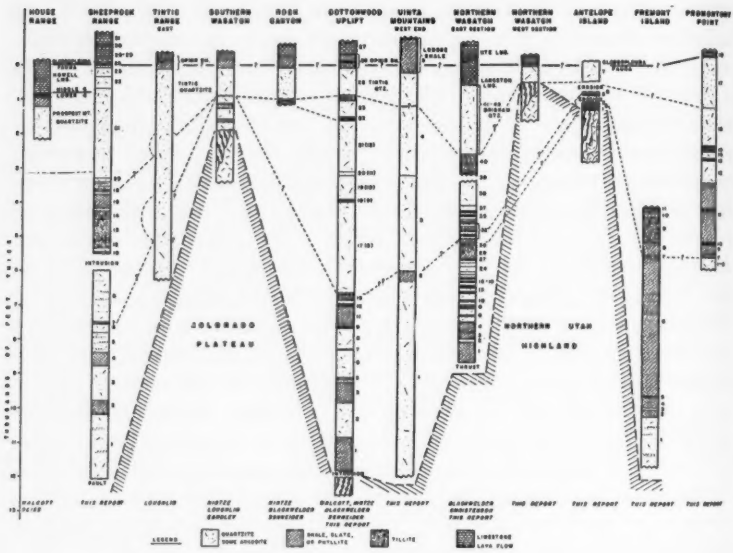

FIGURE 5.—*Correlation chart*

The northern Wasatch east section is a composite of the individual canyon sections of Figure 3. The numbers opposite the columns refer to numbers of units in the text descriptions.

is bottomed on the crystalline complex, but the observation leads the writer to postulate the base of the others not far below the lowest units seen and measured (Fig. 6).

Loughlin and Lindgren (1919, p. 23-25) report 6000 feet of quartzite below the Ophir shale in the Tintic Range. The quartzite is grayish white to pale pink, yellow or brown, fine, even-grained, and composed almost entirely of quartz. Conglomerate beds containing pebbles a quarter of an inch to 2 inches long, chiefly of vein quartz and a few quartzite pebbles, are common throughout the formation. A few thin slate beds were noted. They state that its great thickness is in marked contrast to the Cambrian quartzites elsewhere in Utah and find difficulty in explaining this irregularity. As a result they searched for an unconformity in the quartzite sequence on Quartzite Ridge but found no evidence of one except for

thin conglomerates, supposedly intraformational. They concluded that it must remain an open question as to whether the deposition of sandy sediments began earlier in Cambrian time and progressed much more rapidly or longer in the Tintic district than elsewhere or whether a considerable part of the Tintic quartzite in reality includes strata of pre-Cambrian age, much more purely quartzose than in other sections and essentially identical in character with the Cambrian quartzite.

In making the correlation diagrams (Figs. 5, 6) the latter view of Loughlin and Lindgren is suggested. Is it possible that a small angular unconformity exists at the base of the upper 2000 or 3000 feet of quartzite supposedly of Lower (?) Cambrian age and that units 5 to 8 of the Sheeprock Range are beneath the old erosion surface? This theory would involve the truncation of the argillaceous units 12 to 16 below the unconformity; the two quartzite series would then be in contact with each other (Fig. 6). Lithologically it would be difficult to distinguish the two sequences of quartzites, so no difficulty is met in accounting for the general uniformity of the quartzite sequence in the Tintic district. The theory of two approximately conformable quartzite sequences fits best into the setting of the highland and trough theory that the writers propose later in the article, and in view of the absence of definite supporting evidence of the first theory, *viz.*, continuous conformable series, the second theory is favored.

ORIGIN OF SEDIMENTS

HIGHLANDS AND TROUGHS

The physiographic setting at the time of deposition is of primary concern in formulating a theory of origin of the sediments. It has long been known that the pre-Middle Cambrian strata differ profoundly in thickness from one locality to another in central and northern Utah. After Blackwelder, Hintze, and others located an angular unconformity separating presumably Lower Cambrian quartzites from a thick Algonkian sequence, the theory has been current that at the close of the Proterozoic a revolution of sizable proportions occurred. A uniformly thick and widespread sequence of beds was thrown into great folds; erosion then truncated them in places down to the crystalline complex; and on this surface, considered a peneplain by Hinds (1937, p. 37-41), the Lower and Middle Cambrian sands were spread.

Schneider [4] and Eardley (1939a, p. 1285-1291) questioned certain aspects of this theory when it became apparent through structural studies that a buttress north and northwest of Salt Lake City resisted intense

[4] Hyrum Schneider has postulated a northern Utah highland in lectures at the University of Utah for several years.

deformation during the Laramide revolution while surrounding sedimentary rocks were folded and thrust against it. Isopach maps were made from available stratigraphic data and indicated that the buttress was recurrently positive from Proterozoic (?) time onward.

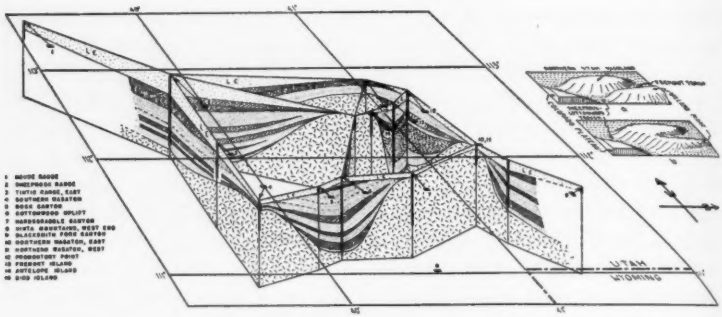

FIGURE 6.—*Perspective correlation diagram*

View is from above the northwest corner of the Colorado Plateau looking westward. The upper datum plane is the Middle Cambrian fossil horizon. It refers to the Tintic and Brigham quartzite formations. The auxiliary diagrams show the two interpretations of the Fremont trough.

The correlation chart of the present report (Fig. 5) strongly suggests the existence of the northern Utah highland, with adjacent deep trough. The fairly reliable correlations in the northern Proterozoic (?) sequences on the basis of the limestones and tillites indicate the thinning and overlapping of the sediments and rather conclusively demonstrate the existence of the highland in Proterozoic (?) time. Hintze (1913, Fig. 4) shows overlapping strata on a northern highland but does not discuss the theory. Because the correlations are three-dimensional in nature, a truer picture of the sedimentary setting is gained by a perspective treatment as worked out in Figure 6. During the Laramide revolution the crust was compressed greatly and the sequences described in the preceding pages were translated several miles from their pre-Laramide positions. The problem of crustal movements in this area has been attacked (Eardley, 1939b, p. 546-552) by the use of a slotted templet. On the basis of this approximate resolution of Laramide movements the former locations of the sections have been plotted in Figure 6. The small black rectangles indicate the present location of sections, and the arrows leading from them point to the approximate pre-Laramide or pre-Cambrian location. This pulling or straightening out is necessary to gain the correct geomorphologic view, particularly in the northern Wasatch, where the thick

offshore sequence has been thrust southwestward over the highland, veneered with only 500 feet of Lower (?) Cambrian beds.

The northern Utah highland appears to be somewhat complex in form inasmuch as a depression or trough in the Fremont Island and Promontory Point area separates it from a distinct ridge along the east side. It is not known whether this Fremont trough extends southward to the Cottonwood uplift because there are no exposures in the intervening area. The two possible interpretations are shown in the small auxiliary diagrams of Figure 6.

The Sheeprock uplift to the south of the northern Utah highland was undoubtedly the site of a trough because sediments in excess of 11,500 feet thick accumulated there. To the east in the southern Wasatch Mountains only 500 to 1000 feet of strata below the unconformity is found. As no correlations can be made between these two sequences it cannot be determined definitely whether the southern Wasatch sequence represents the top of the Sheeprock sequence, and hence the overlapping facies, or whether it represents the bottom of the Sheeprock sequence, and hence what is left after much erosion of a once equally thick sequence. The overlapping theory appears more probable because the Colorado Plateau in the region of the San Rafael Swell, 100 miles to the southeast, was a highland at the beginning of Paleozoic time,[5] and the Uncompahgre highland, 125 miles to the east, was also in existence at the beginning of Middle Cambrian time. The old crystalline complex was probably exposed in each. The central Colorado basin (Lovering, 1933, p. 20-21; Lovering and Burbank, 1933, p. 279-283) was a trough between it and the Front Range highland farther east and during the Paleozoic era it received a great thickness of sediments, due to downwarping of the basin. Especially in Pennsylvanian time narrow troughs, subsiding as much as 20,000 feet, were filled with sediments (T. S. Lovering and K. G. Brill, Jr.: personal communication). The highland and trough theory in north-central Utah during the late pre-Cambrian is, therefore, not without nearby exemplary support.

In view of known highlands in the north-central part of the Colorado Plateau, of stratigraphic evidence of a trough in the Sheeprock uplift area to the west, and of bordering structures, the northwest margin of the Colorado Plateau is thought to have been approximately the northwest margin of a highland during late pre-Cambrian time and is diagrammed as such in Figure 6. (*See also* Figure 5, Eardley, 1939a, p. 1297.)

[5] According to J. M. Kirby of the California Company, Denver, a well was completed in the San Rafael Swell in 1939, which penetrated a granodiorite orthogneiss at 4360 feet. No Proterozoic rocks seem to be present. Such a thin sedimentary veneer was postulated by Baker (1935, p. 1502) some years previously.

ARID FLOODPLAIN THEORY

The presence of ripple marks, mud cracks, and cross-bedding in the argillaceous and arenaceous beds in the Cottonwood uplift, together with the general oxidized iron coloring of the beds, has led Hintze (1913, p. 102) to suggest that the sediments were laid down by rivers on large flood plains in an arid region. This environment, he suggests, would account for the conspicuous dearth of fossils as well as the physical characteristics. Hintze points out, however, that such an environment is not compatible with a glacial theory of origin by which the tillites and varved phyllites are explained. When the thick sequences in all the other localities are examined it immediately becomes apparent that about half the beds of the thick sequences are argillaceous, that red coloring in the argillaceous beds is rare and in the unaltered arenaceous beds somewhat uncommon, and that bedding, except in the tillites, is fairly regular. Even if an arid flood plain were conducive to the formation of red sediments they are present in such limited amounts that such a theory of origin could have only limited application.

GLACIAL THEORY

The tillites were first recognized as such by the late Professor F. J. Pack to whom Hintze (1913, p. 100) credits the discovery and theory. Blackwelder (1932, p. 294, 297) more recently has discussed in some detail the tillites and associated laminated phyllites of the area. He interprets the phyllites as originally varved glacial clays and silts. The dark-gray phyllites that do not exhibit laminations may have been deposited farther from the ice or in interglacial stages or, as suggested by Blackwelder (personal communication), may be marine. Marine and brackish-water clays of Pleistocene age in Finland differ from the lacustrine clays in having no varves. With the measuring of the various sequences two thick tillites were discovered—the 1100-foot unit in the Sheeprock Range and the 700-foot unit on Fremont Island. Thinner tillites on Antelope Island, in Willard Canyon, and below the thicker tillites on Fremont Island and in the Sheeprock Range probably indicate recurrent ice advances. The banded or laminated slates which Blackwelder observed to predominate in the argillaceous units (1932, p. 302) of Little Mountain and Three-Mile Canyons are now known to be present in rather minor amounts in relation to the thick argillaceous units of the major sequences. Hence, the glacio-lacustrine theory of origin cannot apply directly, both areally and stratigraphically, as widely as previously supposed. There is little question, however, about the occurrence of at least one great glaciation. Additional interbedded tillites indi-

cate other recurring glaciations, but their magnitude cannot necessarily be told from the chance thickness of the tillite exposures.

The problem apparently stands as follows: Account for the origin of associated interstratified tillites and varved phyllites on one hand and the quartzites and dark extensive and partly nonlaminated phyllites on the other hand. In view of the highland and trough setting a theory of several closely associated but different sedimentary environments can be visualized, which seems at present the most compatible diagnosis of the known data. It is outlined in the following paragraphs.

MULTIPLE ENVIRONMENT THEORY

Highlands adjacent to subsiding and rather narrow troughs suffered erosion, alternately by glaciers and running water. The highlands were composed chiefly of crystalline, noncalcareous rocks, and mechanical weathering dominated chemical. The weathered debris was deposited in the troughs, thickest in the subsiding axial areas and thinnest in progressively overlapping wedges on the crystalline rock margins of the highlands. Parts of the narrow troughs may have been landlocked at times from invading marine waters, and, if the resulting fresh-water lakes existed at the time of an ice advance, varved clays would have been deposited in them. Marine waters nearby at the same time could have been the site of deposition of dark nonlaminated clays. But the dark nonlaminated clays could also have been deposited in interglacial times in epeiric seas or fresh-water lakes occupying the troughs. Extensive shore lines existed from the beginning to the end of Proterozoic (?) sedimentation. The shores were the site of vigorous wave activity where quartz sands were concentrated and washed clean. No other agency seems capable of producing such great volumes of sand (now quartzite) so widespread, thick, and clean. The sand was supplied by ice and streams of water from the highlands and by wave activity on the highland margins. The waves and shore currents accomplished the sorting and washing and produced much of the cross-bedding visible in the quartzites. The possibility of great deltas whose topset beds were out of water and the site of extensive floodplain deposits of clay, silt, and sand must not be overlooked. The dark coloring of much of the clays and especially the graphitic phyllites in Three-Mile Canyon suggests the existence of aquatic vegetation at the time. A short interval of calcium carbonate deposition occurred on both sides of Willard Ridge. The environmental significance is not clear.

Ice advances, possibly as piedmont glaciers, spread boulder clay over several different kinds of stratified sediment and in one known place—

Antelope Island—on the old crystalline rock. It is possible that much till was reworked by the waves subsequent to deposition and that the glacial climate was of greater duration than is suggested by the relatively thin and widely separated tillite units.

The northern Utah highland and its extension—the Willard Ridge (Fig. 6)—do not appear large enough to have supplied all the sediments that accumulated in the troughs. This leads to the suggestion that the western margin of the Colorado Plateau was the site of a more extensive highland and supplied much of the sediment. Not enough is known about the glaciers to speculate upon their centers of accumulation.

AGE OF SEDIMENTS

SUGGESTED AGES BY PREVIOUS WORKERS

The unconformity 800 feet from the top of Walcott's quartzite section in Big Cottonwood Canyon has been thought to signify a great difference in the age of the quartzite beds above and the quartzite sequence below. The general feeling is expressed in Hintze's words (1913, p. 97):

> "The uniform thickness and wide distribution of the (Cambrian) quartzite and shale member overlying this dividing plane, taken together with the great variation in thickness of the lower series, seem to imply the widespread truncation of the lower beds and their reduction practically to a peneplain before the upper beds were deposited. The complete removal of the great quartzite series over considerable areas must have required much time."

Hintze believes that the Cottonwood quartzites are "Algonkian"—perhaps equivalent to the Belt series of Montana and the Grand Canyon series of Arizona. On the basis of correlation of the tillite with the tillite of the Lower Huronian of Ontario Hintze ventures the suggestion of Lower Huronian age. He (1913, p. 97) also outlines the argument of Daly that, as the lowest fossils are Middle Cambrian, the unconformity may not have the time magnitude suspected, and the Big Cottonwood quartzites may be Lower Cambrian.

Hinds (1936, p. 134) has recently studied and reviewed the Algonkian strata of western North America and concludes that the Big Cottonwood and Uinta quartzites are older than the Beltian rocks and equivalent to the Needle Mountains group of the San Juan Mountains and the Mazatzal quartzite of central Arizona. This correlation would make the Big Cottonwood quartzites middle pre-Cambrian, according to Lovering (personal communication), who hesitates to agree. Hinds' correlation is based chiefly upon his observations of the degree of metamorphism that the various sequences have suffered.

The various bases of age designation indicated in the above references are as follows: the age of the rocks above the unconformity; the amount

of erosion before the deposition of the beds upon the erosion surface; the correlation of the tillites with Lower Huronian tillites of Ontario or with tillites of upper Proterozoic (?) age in Australia and elsewhere; and the difference in degree of metamorphism of the overlying and underlying sequences. Another consideration in point of age is the length of time taken for the accumulation of the entire sequence. Still another is the relation to the Farmington Canyon complex below.

RELATION TO FARMINGTON CANYON COMPLEX

The Farmington Canyon complex (Eardley and Hatch, 1940) is the ancient gneiss and schist foundation upon which the quartzites and phyllites rest. The contact is exposed on Antelope Island, Dry Mountain, and along the Wasatch from Hardscrabble Canyon to Willard Canyon (Figs. 5, 6). The complex is exposed in the mouth of Little Cottonwood Canyon, but the contact is cut out by the Laramide intrusion. A study of the Farmington-Ogden section of the Wasatch reveals the truncation of several thousand feet of the old stratified schists and gneisses. A weathered zone, 3 feet thick, in the schists, is exposed below the contact on Antelope Island, but in the Wasatch the quartzite rests on slightly weathered crystalline rocks. It is evident that the absence of a weathered zone does not signify a short period of erosion. On the other hand a long time interval is represented by the unconformity, in view of the profound difference in metamorphism of the two series. The lower has been extensively intruded, injected, and dynamically metamorphosed. The intrusions and injections and most of the dynamic metamorphism preceded the deposition of the overlying sediments. The extensive erosional beveling of the gneisses, schists, and granites before the deposition of the younger rocks also signifies considerable age difference.

Lovering (1935, p. 13) has had lead, uranium, and thorium determined for the Pikes Peak granite of the Front Range of Colorado and finds the analyses to signify an early middle Algonkian age for the rocks. The Pikes Peak granite is a late stage of the batholithic cycle there. After observing hand specimens of the northern Utah crystalline rocks and reading this report in manuscript form, he recognizes a stage of metamorphism similar to that of the earliest pre-Cambrian rocks in the Front Range. These earliest rocks belong to the beginning phases of the batholithic cycle. On this sketchy basis the rocks described in the foregoing pages are later in age than middle pre-Cambrian.

RELATION TO CAMBRIAN FORMATIONS

The faunal zones of all the Cambrian formations in Utah have not been completely studied, and, as a result, the problem arises as to whether

the barren quartzites below the recognized Middle Cambrian shales and limestones are Middle or Lower Cambrian.

Gilluly (1932, p. 19, fig. 3) compiled a chart of the Cambrian formations of Utah and Nevada and, on the basis of existing paleontologic information, correlated them along the lowest established Middle Cambrian beds. According to Deiss' publication (1938, p. 1161-1164; fig. 7) and according to present beliefs expressed by him in personal discussion this datum plane is approximately the *Glossopleura-Kootenai* and recurrent *Glossopleura* faunal zones. The *Glossopleura* fauna is common in the Ophir shale, particularly near the base. In the construction of the chart (Fig. 5) of this report the *Glossopleura* fauna insofar as known was used as the datum plane. Cambrian fossils older than the *Glossopleura* fauna have been recognized in Utah only from the House Range (Deiss, 1938, p. 1141, fig. 5). Resser (1939, p. 6) very recently states that Lower Cambrian seas invaded the region from the west in the southern Wasatch, in the Cottonwood uplift, and on Promontory Point, but no paleontologic information is given, however, to substantiate the conclusion. If Lower Cambrian fossils are found, they will probably come from the basal beds of the Ophir shale or from shale partings between quartzite beds in the Tintic and Brigham quartzites.

A tripartite division (limestone, shale, quartzite) of the Cambrian beds exists in the East Tintic Range, in the Ophir district, and in the Wasatch Range from Dry Mountain to Willard Canyon. Each division is lithologically similar and fairly uniform in thickness, except the thick quartzite in the East Tintic Range. It is unfortunate but unavoidable that this section where the base is questionably defined is the type locality of the Tintic quartzite.

The type locality of the Brigham quartzite is in the Wasatch front northeast of Brigham City (Resser, 1939, p. 12; Deiss, 1938, p. 1118; Walcott, 1908a, p. 8). This quartzite is more than twice as thick as the Tintic (?) quartzite 6 miles to the south in the mouth of Willard Canyon. It rests on the thick phyllite and quartzite sequence of the overthrust sheet, whereas the Tintic quartzite of Willard Canyon rests upon the ancient crystalline complex and is believed to be the overlapping equivalent of the Brigham quartzite which accumulated, perhaps under the same environmental conditions, near the central part of the subsiding trough. The Brigham quartzite of the Bear River Range 30 miles east of the type locality is properly correlated and, according to Resser (1939, p. 12-19), is overlain by beds containing the *Ptarmigania* fauna which is Middle Cambrian. According to Deiss, the *Glossopleura* zone of this fauna can be correlated with that in the Ophir formation which overlies the Tintic quartzite.

From existing paleontological work it may be concluded that the Brigham and Tintic quartzites are approximately the same age and pre-date the Glossopleura fauna which is somewhat older than middle Middle Cambrian. From tectonic data the Tintic quartzite in the northern Wasatch Mountains is the overlapping shore equivalent of the Brigham quartzite, and hence slightly younger. Where separated by an observed unconformity from the strata below, their thicknesses range from 500 to 1000 feet, and it seems probable that their lower beds date back to at least early Middle or late Lower Cambrian time. Much of Lower Cambrian time does not appear to be represented by the quartzite beds above the unconformity, and upon this basis the deformation and erosion indicated by the unconformity might have occurred entirely in Lower Cambrian time.

The significance of the unconformity has been discussed previously in this article and elsewhere (Eardley, 1939a). According to the highland and trough theory the deformation was not of great diastrophic proportions, and subsequent erosion was much less than Hintze (1913, p. 97) visualized. Over 10,000 feet of strata was previously thought to have been eroded; it is now believed that the thickness was not more than 3000 feet—probably less in most places. Moreover, as no unconformity has yet been found in most of the central trough sequences where the beds are concordant or nearly so, it is possible that the upper several thousand feet, at least, of the pre-Ophir strata is Lower Cambrian.

AGE SIGNIFICANCE OF MULTIPLE ENVIRONMENT THEORY

If repeated glaciations occurred during the accumulation of a large part of the trough sequences then it might be inferred that a time no longer than the Pleistocene epoch is represented. This could conveniently be placed in the Lower Cambrian, and the further inference be drawn that no part of the thick sequences is Proterozoic. Recurring volcanism is believed to be recognized in the northern areas, again suggesting an episode of volcanism during the accumulation of 3000 to 4000 feet of the sediments, and a relatively short period of time.

Narrow basins along the Pacific in California subsided in Pliocene or Pleistocene time to depths greater than any of the troughs of the northern Utah area. The great thickness of the trough sequences cannot, therefore, be safely taken to indicate great time in accumulation.

METAMORPHISM OF QUARTZITES

Evidence assembled thus far does not indicate great diastrophic intensity in the disturbance resulting in the unconformity. However, a method of approach to this problem not yet mentioned is field and

petrographic observation of metamorphic effects in the quartzites above and below the unconformity.

Megascopically the repeated observation in all sequences on the basis of hardness and quartzitic appearance point to the conclusion that the quartzites below the lowest Middle Cambrian fossil zone, whether above or below the unconformity, cannot be safely distinguished from each other. Sandstones and quartzitic sandstones occur thousands of feet below the fossils, or below the unconformity, and very hard quartzites occur above. In fact, as Professor Schneider has pointed out to his classes for many years, quartzite beds occur in several of the overlying Paleozoic and Mesozoic formations and commonly are interbedded in sandstones. The quartzitic nature probably depends locally and chiefly on the degree of siliceous cementation and not on the degree of metamorphism.

Under the microscope the few slides studied to date indicate that there are two types of quartzites in the post-Farmington Canyon complex rocks and a third type within the ancient complex.

The Cambrian quartzites and quartzitic sandstones are sandstones with silica cement. The silica has been added to the original rounded sand grains in optical continuity. A thin iron oxide border around most of the grains enables one to distinguish the original grains from the added material. The size of the sand grains varies from about a quarter of a millimeter to 1 millimeter in diameter. Slight to moderate straining of the quartz grains and cement is revealed by the undulatory extinction. Other minerals beside quartz constitute less than one per cent of the rock and include small inclusions of magnetite, zircon, rutile, and apatite in quartz. Insufficient microscopic work has been done to substantiate the conclusion that this type of quartzite is the most abundant by far of all the arenaceous rocks above and below the unconformity, but field observations lead the writer tentatively to such a conclusion.

The sugary-appearing metaquartzites from the lower part of the sequence in Willard Canyon have been metamorphosed. The grain size varies from about one-twentieth to three-quarters of a millimeter in diameter. Many of the larger grains are about twice as long as they are wide, the elongations tending to be parallel with one another and thereby imparting a lineation in the rock which is visible even in the hand specimen. The small grain size and the relation of small grains to the larger suggest that the rock was granulated during metamorphism. Subsequent recrystallization, probably during the same period of metamorphism, has healed the grains together and produced sutured contacts particularly noticeable between the larger grains. The quartz shows strong undulatory extinction. Small needles of muscovite constitute about

one per cent of the rock. Minute quantities of magnetite, zircon, and apatite occur as inclusions in the quartz.

Blackwelder (personal communication) has suggested that the sugary metaquartzites and interbedded slates, phyllites, and schist may represent a sequence of intermediate age between the Farmington Canyon complex and the beds generally regarded as Proterozoic. Special observations bearing on this point are as follows:

The argillaceous beds in the lower part of the Willard Canyon sequence are advanced in degree of metamorphism in approximate proportion to the thickness of the inclosing quartzite beds and inversely to their own thickness, *viz.*, if the quartzite series above and below are each several hundred feet thick, the separating argillaceous beds if also several hundred feet thick will be phyllitic; if the quartzites are equally as thick, and the argillaceous layer only 10 feet thick, the argillaceous beds will be schistose. (*See* unit 12 of Willard Canyon.) The presence of beds of schist in this sequence does not signify an older age but a different metamorphism local to the beds and to the area. The sequence is terminated below by the Willard thrust, and dynamic metamorphism of Laramide age might have been especially severe to beds at the base of the sequence inasmuch as the overriding thrust sheets met resistance in the northern Utah highland. Nothing was found in the lithology or stratigraphic sequence of the beds to suggest that they are a distinct group from the quartzite and phyllite sequences elsewhere in the region.

In contrast to the above two types of quartzite, that from the Farmington Canyon complex is strongly sheared and granulated. Grain size varies from quartz flour in the shear zones to over a millimeter in length. Some of the larger grains are pulled out until their length is five to ten times their width. The quartz is strongly strained. There has been considerable recrystallization of the quartz in order to evolve the exceedingly elongate grains. However, the cataclastic texture is very conspicuous. Large patches of a sericitic aggregate are found in some of the quartzites and probably represent former feldspar grains. Very finely granular sericite is commonly strung out in the shear zones. Small amounts of the usual accessory minerals are present. It is evident that this metaquartzite has suffered much more intense metamorphism than the quartzites of the overlying sequences and that a period of strong crustal deformation preceded the accumulation of the overlying series.

A petrographic and petrofabric study of suites of oriented quartzite specimens from top to bottom of the various sequences is contemplated. Before this is finished, it would be unsafe to draw further conclusions.

CONCLUSIONS

The trough sequences of north-central Utah are considerably younger than the underlying Farmington Canyon complex which is about middle pre-Cambrian. They are older than the *Glossopleura* fauna which is somewhat older than middle Middle Cambrian. Although the sequences in places are separated from quartzites underlying the *Glossopleura* fauna by a 5- to 15-degree angular unconformity, they appear elsewhere to be conformable and continuous. The unconformity probably signifies uplift and erosion chiefly around the highland margins and does not imply folding and erosion on a great scale as previously theorized. Study of the beds above and below the unconformity does not indicate that the disturbance resulting in the unconformity was severe enough to produce any metamorphic effects. Recurrent glaciations through a great part of the strata and an episode of volcanism during the accumulation of 4000 feet suggest that the sequences were deposited during a comparatively short time. It is quite possible, therefore, that the beds below the unconformity are Lower Cambrian. It is also possible that only the upper part of the phyllites and quartzites are Lower Cambrian and that the lower part is upper Proterozoic.

As pointed out by Howell (1937, p. 208), the tillites may have been deposited when a glacial climate engrossed the world. Tillites in southern Australia, central and western China, Central Africa, in the Torridonian of Scotland, southern Norway, Spitzbergen, east Greenland, and in the Keweenawan series of Lake Superior are supposedly upper Proterozoic or Lower Cambrian and may be viewed according to this theory as of very similar age. For want of any better marker of the close of the Proterozoic era it may be advisable eventually, according to Howell, to draw an arbitrary division between Cambrian and Proterozoic at the top of the tillites representing the glacial climate.

In view of the above age indications it does not seem possible to accept Hinds' (1936, p. 134) theory that the beds are pre-Beltian.

SUMMARY OF CONCLUSIONS

It is concluded that sediments accumulated in subsiding troughs to thicknesses of more than 12,000 feet. The troughs were in part narrow and long. One trough called the Sheeprock-Cottonwood apparently extended northeastward from the Sheeprock Range to the Cottonwood uplift and hence eastward along the site of the Uinta Mountains. It was bordered on the northwest by the northern Utah highland and on the southwest and south by a highland along the margin of the Colorado Plateau. A high ridge as part of the northern Utah highland extended

northward from Salt Lake City. On its west side was the Fremont trough and on its east was the site of deposition of beds in the Willard overthrust sheets and in the Bear River Range.

The crystalline highlands were vigorously eroded by glaciers and running water. Mechanical weathering dominated. The sediments accumulated thickest in the centers of the troughs and in over-lapping fashion on the highlands. The troughs may have been landlocked at times from invading seas, and in fresh-water lakes existing at the time of recurrent glaciations varved clays and silts were deposited. Marine waters nearby in the troughs were the medium of deposition of dark nonlaminated or poorly laminated clays. Extensive shore lines existed from the beginning to the end of trough filling and were the site of vigorous wave activity. Great volumes of sand were here concentrated and washed clean. Large deltas were built in the troughs whose tops were often out of water and the site of extensive clay, silt, and sand floodplain deposits. The dark coloring of many of the argillaceous beds, especially the graphitic phyllites in Three-Mile Canyon, suggests the existence of aquatic plants at the time. A short interval of calcium carbonate precipitation occurred on either side of Willard Ridge.

Ice advances, possibly as piedmont glaciers, spread boulder clay over several different kinds of stratified sediment, and in one place, Antelope Island, on the old crystalline rocks. The sediments were deposited in late Proterozoic or early Cambrian time, or during an interval that transgressed both. They were then tilted up along the highland margins, beveled, and covered by sands of the Tintic and Brigham quartzites. These quartzites above the unconformity are earliest Middle Cambrian or late Lower Cambrian. In the central trough areas, except in the Cottonwood uplift, no break between the Brigham or Tintic quartzite and the lower quartzites and phyllites has been found.

The variable quartzitic nature of the arenaceous sediments is chiefly due to differential silica cementing. No difference in degree of metamorphism of the quartzites immediately above or below the unconformity could be discerned. The metamorphism of the argillaceous sediments is believed to have occurred in large part during the Laramide revolution while they were buried under a cover of strata 20,000 to 30,000 feet thick.

WORKS TO WHICH REFERENCE IS MADE

Baker, A. A. (1935) *Geologic structure of southeastern Utah,* Am. Assoc. Petrol. Geol., Bull., vol. 19, p. 1472-1507.

Blackwelder, Eliot (1910) *New light on the geology of the Wasatch Mountains, Utah,* Geol. Soc. Am., Bull., vol. 21, p. 517-542.

Blackwelder, Eliot (1932) *An ancient glacial formation in Utah,* Jour. Geol., vol. 40, p. 289-304.

——— (1935) *Summary of the pre-Cambrian rocks of Utah and Wyoming,* Utah Acad. Sci., vol. 12, p. 153-157.

Butler, B. S., *et al.* (1920) *Ore deposits of Utah,* U. S. Geol. Survey, Prof. Paper 111, 672 pages.

Deiss, Charles (1938) *Cambrian formations and sections in part of Cordilleran trough,* Geol. Soc. Am., Bull., vol. 49, p. 1067-1168.

Eardley, A. J. (1932) *Stratigraphy of the southern Wasatch Mountains,* Mich. Acad. Sci., Papers, vol. 18, p. 307-344.

——— (1939a) *Structure of the Wasatch-Great Basin region,* Geol. Soc. Am., Bull., vol. 50, p. 1277-1310.

——— (1939b) *Slotted templet for resolving crustal movements,* Jour. Geol., vol. 47, p. 546-554.

———, and Hatch, R. A. (1940) *Pre-Cambrian crystalline rocks of north-central Utah,* Jour. Geol., vol. 48, p. 58-72.

Emmons, S. F. (1890) *Orogenic movements in the Rocky Mountains,* Geol. Soc. Am., Bull., vol. 1, p. 258.

Forrester, J. D. (1937) *Structure of the Uinta Mountains,* Geol. Soc. Am., Bull., vol. 48, p. 631-666.

Gilluly, James (1932) *Geology and ore deposits of the Stockton and Fairfield quadrangles, Utah,* U. S. Geol. Survey, Prof. Paper 173, 167 pages.

Hinds, N. E. A. (1935) *Ep-Archean and ep-Algonkian intervals in western North America,* Carnegie Instit. Washington, Pub. 463, p. 1-52.

——— (1936) *Uncompahgran and Beltian deposits in western North America,* Carnegie Instit. Washington, Pub. 463, p. 53-136.

Hintze, F. F., Jr. (1913) *A contribution to the geology of the Wasatch Mountains,* N. Y. Acad. Sci., Ann., vol. 23, p. 85-143.

——— (1934) *The Proterozoic-Paleozoic contact in the western Uinta and central Wasatch Mountains, Utah,* Utah Acad. Sci., vol. 11, p. 165-166.

Howell, B. F. (1937) *Late Proterozoic and early Cambrian climates,* 17th Intern. Geol. Congr., Russia, Abstracts of papers, p. 208.

King, Clarence, Emmons, S. F., and Hague, Arnold (1877) *Report of the United States Geological Exploration of the 40th Parallel,* vols. 1, 2, 803 pages.

Lovering, T. S. (1933) *Colorado,* Intern. Geol. Congr., Guidebook 19, 146 pages.

——— (1935) *Geology and ore deposits of the Montezuma quadrangle, Colorado,* U. S. Geol. Survey, Prof. Paper 178, 119 pages.

———, and Burkank, W. S. (1933) *Ore deposits of the western United States* (Lindgren volume), Am. Instit. Min. Metall. Eng., p. 272-315.

Lindgren, Waldemar, and Loughlin, G. F. (1919) *Geology and ore deposits of the Tintic mining district, Utah,* U. S. Geol. Survey, Prof. Paper 107, 282 pages.

Powell, J. W. (1876) *Report on the geology of the eastern portion of the Uinta Mountains and a region of country adjacent thereto,* U. S. Geol. Geog. Survey Terr., II Div., 218 pages.

Resser, C. E. (1939) *The Ptarmigan strata of the northern Wasatch Mountains,* Smithson. Misc. Coll., vol. 98, no. 24, 72 pages.

Schneider, Hyrum (1925) *A discussion of certain geologic features of the Wasatch Mountains,* Jour. Geol., vol. 33, p. 28-48.

Stansbury, Howard (1853) *Exploration and survey of the valley of the Great Salt Lake of Utah,* U. S. 32d Congr. Spec. Sess., Sen. Ex. Doc., 487 pages.

Walcott, C. D. (1886) *Second contribution to the studies on Cambrian faunas of North America,* U. S. Geol. Survey, Bull. 30, p. 38-39.

—— (1891) *Cambrian correlation papers,* U. S. Geol. Survey, Bull. 81, p. 313-326.

—— (1908) *Nomenclature of some Cambrian Cordilleran formations,* Smithson. Misc. Coll., vol. 53, no. 5, p. 167-230.

Weeks, F. B. (1907) *Stratigraphy and structure of the Uinta Range,* Geol. Soc. Am., Bull., vol. 18, p. 444.

Winchell, A. N. (1936) *A third study of chlorite,* Am. Mineral., vol. 21, p. 642-651.

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CRUSTAL SHORTENING AND AXIAL DIVERGENCE IN THE APPALACHIANS OF SOUTHEASTERN PENNSYLVANIA AND MARYLAND

BY ERNST CLOOS

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ABSTRACT

Calculation of crustal shortening ratios in the northern Appalachians has been undertaken on the basis of continuous exposures provided by the anthracite fields of Pennsylvania. The ratios increase from northeast to southwest with a maximum

approximately in the Harrisburg section. The increase is accompanied by convergence of axes from the northeast toward the Susquehanna River, and consequently the Appalachian trends bulge northwestward. An arc is formed with Baltimore approximately at the center of curvature. In the same measure as the folds crowd northwestward thrusts appear in the crystalline axis and involve the Glenarm series and the lower Paleozoic rocks. The gneiss basement does not participate with equal intensity in the folding, and an axial divergence between basement and cover becomes visible. Most of the "Appalachian folding" seems to have taken place above the gneiss complex and the basement is only superficially involved. The Catoctin volcanic complex and the Glenarm series participate in the Appalachian trend. This seems to indicate a major structural break above the Baltimore gneiss complex.

INTRODUCTION

Attempts to calculate the amount of shortening of the earth's crust due to folding are numerous. R. T. Chamberlin (1910) measured approximately 400 attitudes along a section from Harrisburg to Tyrone, Pa. (Fig. 3). He then calculated the amount of shortening from the section constructed. In a later paper (1919) he estimated the shortening in the Rocky Mountains. Van Hise (1898) estimated the amount of shortening on the basis of more general considerations. Albert Heim's analysis (1919) of crustal shortening is well known. Mansfield (1927) discussed lateral shortening in the Rocky Mountains in some detail. However, his section is unfortunately broken by thrusts and faults, the lateral components of which are only estimated. Recently P. B. King (1937, p. 133) estimated shortening in the Marathon region in Texas.

In the following the author has attempted a calculation over a larger area in eastern Pennsylvania on the basis of sections not used before. He then applies the calculation in an attempt to interpret the large northwestward bulge of the Appalachian folds and discusses some of the effects of the swerving axes and their relationship to the structures in the basement complex.

The author does not intend to present dogmatic theory but submits working hypotheses and some broader views which deserve attention. Problems are presented, not solutions. Some of the more urgent detailed investigations have been undertaken and are either completed (Broedel 1937; Hershey 1937; Marshall 1937; Cohen 1937; Cloos and Broedel 1940) or are near completion (Cloos, Hietanen, and Guild 1940; Mather 1940; Broughton 1940); others will soon be started.

Fortunately an important part of the area is covered by excellent reconnaissance maps of the Maryland and Pennsylvania Geological Surveys, by Florence Bascom, G. W. Stose, A. I. Jonas, and E. B. Knopf to whom all later workers owe a great debt of gratitude.

METHOD OF CALCULATION

Any natural fold shows a determinable amount of horizontal relative movement of two parts of a stratum toward each other. Additional movements in other directions may occur; if the path of a sand grain

in a sandstone of the Alps or Appalachians could be traced back to the point of deposition it would be a highly complicated curve, and not a simple horizontal line from one point to the other within one plane. In

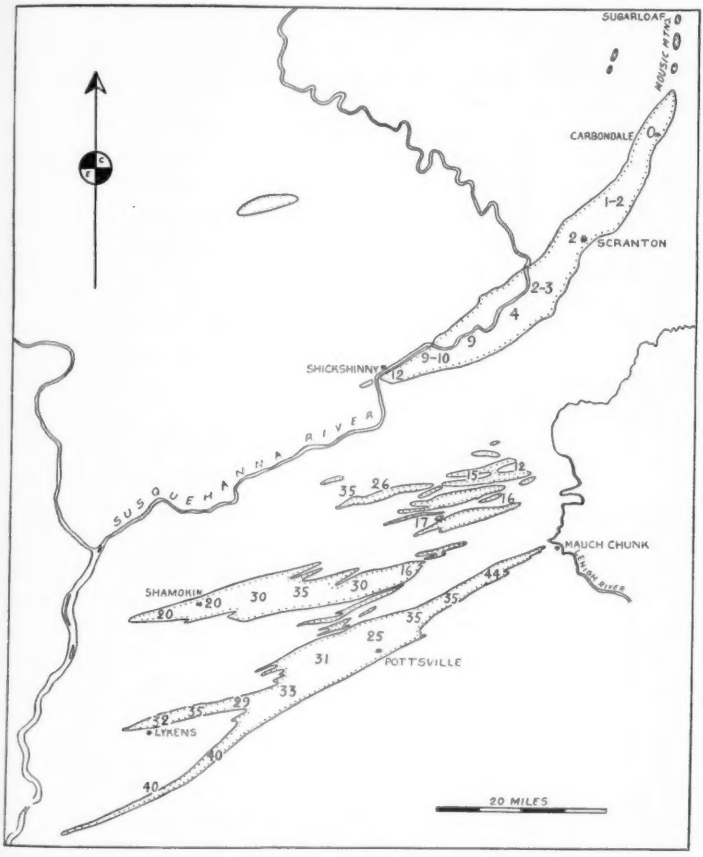

FIGURE 1.—*Percentage of shortening in the anthracite coal fields of Pennsylvania*

spite of the complicated nature of such a movement, it is possible to determine at least the horizontal component of such paths as the difference between a formerly horizontal bed and the present crumpled or folded condition of the same bed. The values obtained do not include vertical components of any kind, or repetition of movements over the same or similar paths.

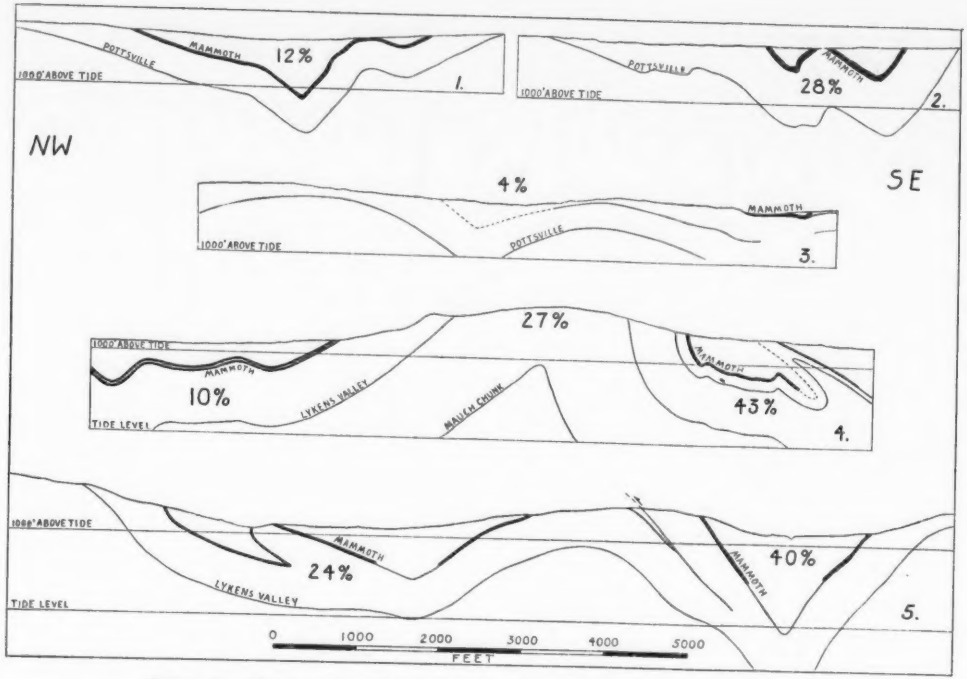

FIGURE 2.—*Five sample sections through the anthracite area of Pennsylvania*
Vertical equals horizontal scale. Copied from sections in the reports of the Second Geological Survey of Pa.

In order to determine the horizontal component one should measure the distance "a" between two points along the bedding and then measure the present horizontal distance "b" between the two end points. The value $100 - \frac{b}{a} \cdot 100$ indicates the amount of shortening in percent. The horizontal approach of the two points toward each other is a-b (Heim, 1919, p. 649).

These values, however, give no information about lateral compression, direction of movements, or anything but shortening in a stratum due to deformation. In schistose rocks the value cannot be determined as long as the lateral component of the movements on millions of cleavage planes cannot be determined. Metamorphic rocks of all kinds are excluded because volume changes due to recrystallization cannot be estimated and frequently bedding is obliterated by shearing. If faults or thrusts transect a chosen section it can be used only if the exposures are complete and the horizontal component of such displacements can be measured. Unfortunately, there are so many restrictions to such calculations that only very few areas can be used.

The two measurements can be made with an ordinary scaling wheel thus including the same errors in both measurements and providing the length of the section directly.

ANTHRACITE AREA IN EASTERN PENNSYLVANIA

Completely exposed sections are found in areas in which mining operations have provided continuous exposures. In a careful search of the literature the writer found that the reports of the Second Geological Survey of Pennsylvania on the anthracite area are particularly satisfactory for the purpose. Large scale maps and sections without vertical exaggeration are abundant (Second Geological Survey, 1886).

The author used these reports for the following calculations and considerations. One hundred forty-three sections have been measured and the values tabulated in Table 1. Figure 1 shows the average values plotted on a map of the area.

Five typical sections in Figure 2 show clearly that short sections or portions of sections are misleading. A minimum length has to be reached and all possible flexures must be included. A comparison of values in Table 1 shows that such a minimum may be near 2500 feet in the area under discussion. Longer sections show lower values since local and very intense disturbances are compensated by broad anticlines and synclines. A similar observation can be made by comparison of Heim's (1919, p. 650) calculations of crustal shortening in the Jura mountains.

Faults and thrusts have to be eliminated and calculated separately and can be added or subtracted later.

Vertical equals horizontal scale. Copied from sections in the reports of the Second Geological Survey of Pa.

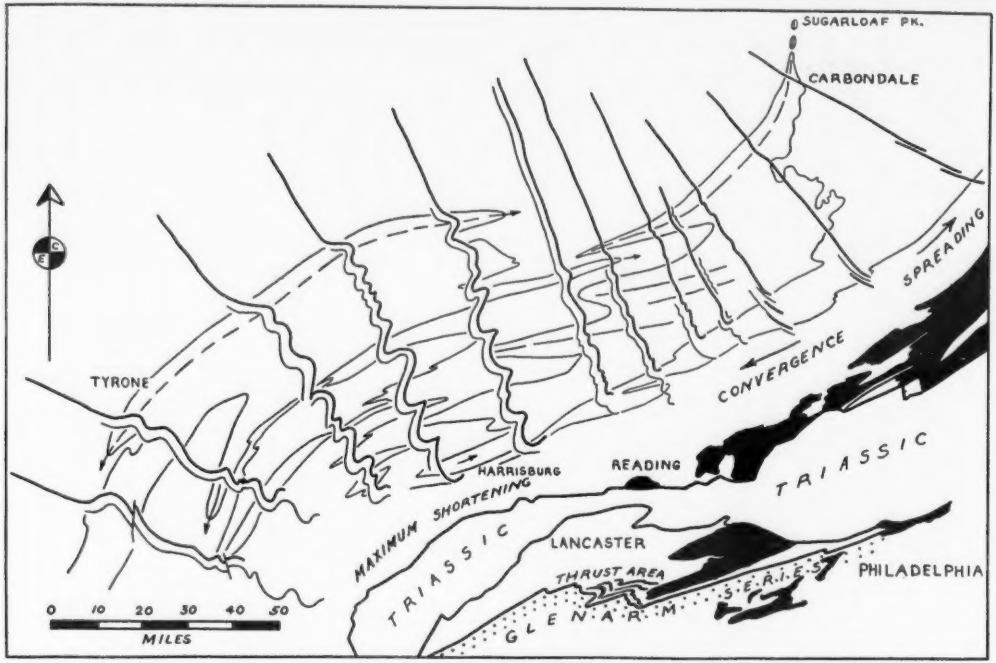

FIGURE 3.—*Generalized sections across the northern Appalachians in eastern Pennsylvania*

Sections are drawn on the Oriskany, but are designed to show only schematically the intensity of folding and the number of folds present. Solid black: Baltimore gneiss complex.

In Figure 2 the "Mammoth bed" is indicated as reference layer. Below the "Mammoth bed" the Pottsville conglomerate or the "Lykens Valley" bed are marked to indicate the thickness of the section. The vertical equals the horizontal scale in all sections. Sections 1 and 2 are very similar and show a surprisingly low value for each syncline. The length of the sections does not enter the ratio directly. Two folds in section 2 show 28 percent shortening, but one fold of twice that length in section 1 shows only 12 percent. Section 3 has been included to demonstrate why some of the data are unreliable. In this profile the "Mammoth bed" is exposed only for a short distance and a construction of the Pottsville seems to indicate only 4 percent shortening. As a rule, generalized sections have lower values of shortening because all small folds have been lost in the generalization. Section 4 consists of three parts with widely different values. The northwestern portion includes gently warped beds which are shortened only 10 percent. The southeastern part is an overturned fold, possibly with a thrust; that portion alone has been shortened 43 percent. In section 5 the same amount is obtained in one simple but steep syncline. The northwestern portion of that section shows less reduction than should be expected because of the large open syncline. No average has been calculated for section 5 because of the probable thrust between the two parts.

The ratios plotted in Figure 1 represent average values for each surrounding area. As far as possible only sections over one mile long have been used since generalizations from intense local disturbances would exaggerate the values.

GEOGRAPHICAL DISTRIBUTION OF SHORTENING RATIOS

In eastern Pennsylvania crustal shortening increases parallel to the Appalachian trend from northeast to southwest as well as across the strike from northwest to southeast. The latter is well known and frequently mentioned as an increase in the intensity of folding. Most important is the increase within the northern anthracite field between Carbondale and Shickshinny (Fig. 1). At Carbondale and northward folding is so slight that lateral shortening cannot be measured. The northernmost portion of the syncline is exposed in Sugarloaf Peak, a remnant of Pottsville sandstone which appears very conspicuous on the geologic map of Susquehanna and Wayne counties by I. C. White (1881). From Carbondale to the tip of the syncline the axis strikes north northeast and then north-south in the Mousic Mountains to Ararat and Sugarloaf Peaks. From Carbondale to Shickshinny the ratio increases from zero to 12 percent. Simultaneously the synclinal axis changes its strike from N.–S. to N. 65°–75° E. Proceeding toward the southwest the amount

number of folds present. Solid black: Baltimore gneiss complex.

of shortening increases gradually to over 40 percent in the southern Anthracite fields. The east and west terminations of the Pottsville area show a high shortening ratio of over 40 percent. This value is probably

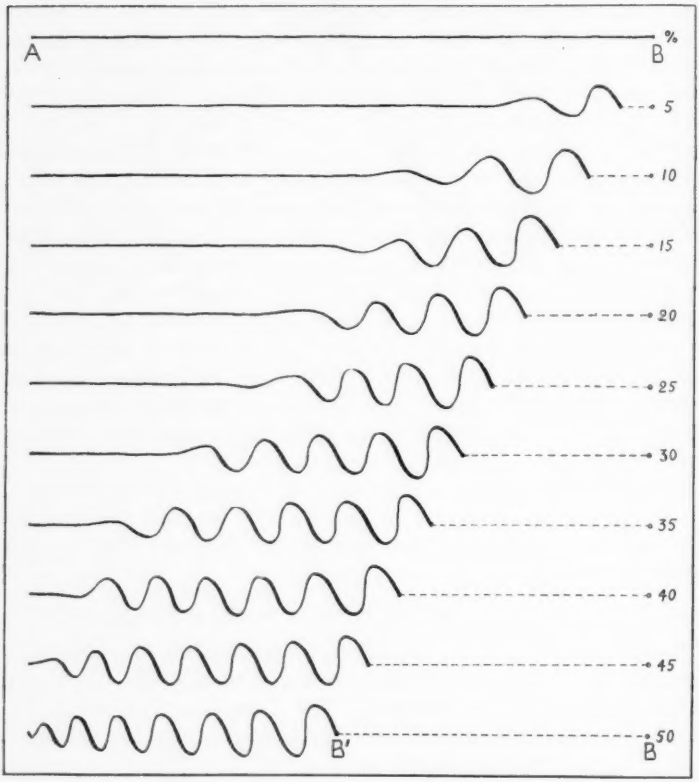

FIGURE 4.—*Schematic diagram drawn from bent wire to indicate the movement of point B toward B' due to folding*

Percentage of shortening and number of folds present are shown.

exaggerated and seems to be due to the overemphasis of one simple syncline (Fig. 2, section 5) in a short section. At Mauch Chunk the maximum reduction is 50 percent but here also only one fold has been exposed and measured. The southwestern end is similar and also not reliable. The average value of 30 to 35 percent near Pottsville is probably the most accurate, since it has been calculated from long sections.

AXIAL DIVERGENCE AND LATERAL SHORTENING

As a logical consequence of the dying out of folds the distance between anticlinal or synclinal axes increases. If the amount of lateral shortening is reduced the axes diverge and if the ratio becomes zero the folds may change their trends considerably.

This fact is rarely mentioned in the literature. Albert Heim (1919, p. 630-31, fig. 111) describes the dying out of a fold:

"Wenn der eine Teil eines gespaltenen Gewölbes rasch abnimmt, ohne durch ein anderes Gewölbe daneben ersetzt zu werden, so bedingt dies in freierem Terrain fast immer eine Umbiegung der erlöschenden Kette nach der Innenseite des Faltenbogens." (If the one half of a double anticline dies out rapidly without being replaced by another one next to it, the anticline must bend towards the inside of the fold-arc if the terrane is sufficiently open.—Free translation by E. C.)

Assuming the distance between two major folds as one mile and a reduction of the amount of lateral shortening from 50 percent to zero along the strike, the distance between the two axes increases to 2 miles. This increase necessitates a change of the axial strike and the abruptness of this change depends on the rate of the decrease of the ratio of lateral shortening along the strike.

In the Appalachians of eastern Pennsylvania the shortening ratio above the Juniata sandstone decreases from 30-35 percent in the heart of the anthracite region to zero near Carbondale in the area north of the Delaware River. At the same time the axes in the northern anthracite fields (Shickshinny-Carbondale axis Fig. 1) swerve from N. 60° E. to N.–S.; other axes spread accordingly. The distance between the Carbondale axis and the Blue Ridge extension along the upper Delaware River anticline (Fig. 3) increases from 28 to 42 miles as measured from the exposure of the Juniata sandstone to the center of the axis, or about 33 percent. This is the same amount of shortening as determined in the analysis of over 100 detailed sections and compares very well with the figures plotted in Figure 1. The same effect is illustrated in Figure 3 which shows, schematically, the amount of axial divergence and its relation to the intensity of folding. The diagram, Figure 4, has been constructed from a strip of wire, A-B, which has been bent into small folds step by step. The maximum shortening is 50 percent, and point B has migrated from its original position to B′.

AXIAL CONVERGENCE AND CONVEXITY

Reversing the above outlined consideration one might consider the convergence of the axes from northeast to southwest. Proceeding from the Adirondacks southwestward a number of folds appear in eastern Pennsylvania converging towards Harrisburg and the Susquehanna River.

The increase in the intensity of folding is shown in an increase of the ratio of lateral shortening (Fig. 1). In Figure 3 twelve sections have been plotted across the Appalachian Mountains between the upper Delaware River (Catskill area) to the Maryland line. The sections are generalized because of lack of accurate data. In those regions where new maps are not yet available by far the best sources of information are the reports and maps of the Second Geological Survey of Pennsylvania.

Within the anthracite area the average lateral shortening reaches 35 percent (locally 44 percent) and increases probably towards the southeast with the appearance of thrusts.

Proceeding from the Catskill Mountains toward the southwest the Carbondale syncline swerves from north-south to southwest, and in its projected extension an additional group of folds appears west of the Susquehanna River. The total amount of shortening increases with the addition of these folds and because of this increase the Appalachian folds bulge northwestward.

Figure 4 demonstrates the reason for convex bulging. With increasing shortening and an added number of folds, point B finally migrates to B′ if A is kept stationary. The amount of migration of B depends entirely on the intensity of folding at any given locality and varies accordingly.

Proceeding southward from the Susquehanna River into Maryland the axes swerve back to almost north-south and parallel to the northern portion.

The best analogy to the northern Appalachians is found in the Swiss-French Jura Mountains (Bucher 1933, p. 240). There the strike changes from east-west at the east end south of the Black Forest to north-south west of Geneva, between the Alps and the Central Plateau (Fig. 6). The greatest total intensity of folding is found in the maximum bulge. Here the largest number of folds appear (Heim 1919, p. 648-650) and the maximum migration of B toward B′ (Fig. 4). The curvature seems to be a function of the amount of lateral shortening of the folded strata. It is not at all necessary that such an area coincide with the area of maximum thrusting or the most intense local folding. Ten large, open folds (Fig. 2, sec. 5) are the result of more lateral shortening than one or two thrusts (Fig. 2, sec. 4). The intense overturning and thrusting south of the Black Forest neither indicates the maximum amount of shortening, nor the greatest intensity of folding.

The writer does not attribute all arcuate mountains to this cause. He believes, however, that the northern Appalachians and the Jura arcs are a function of the number of folds and the total intensity of folding in

each section. The maximum amount of shortening seems to be located in the area with the maximum bulge. At the same time the arcuate shape of the present mountain system does not mean that an arcuate geosyncline preceded it; it may well be that Ordovician organisms now unearthed near York, Pa., died in the vicinity of Baltimore, Md.

"APPALACHIAN" TRENDS

"Appalachian" trends are those directions which are the result of Appalachian folding. Several periods of deformation may be hidden within these structures and older trends may have preceded the later ones. The most obvious, however, are the "Appalachian" trends which appear on the geologic maps of Pennsylvania (Stose, 1933) and on the geologic map of the United States (1933). This trend is almost north-south in New York; it bends to the southwest and strikes almost east-west in eastern Pennsylvania, then turns southward again and becomes almost north-south in Maryland before the Potomac River is reached.

Seemingly all formations participate in this great arc of which Baltimore is the approximate center. The axes of all folds comprising all sediments from the Cambrian to the upper Pennsylvanian participate in it and large portions of the pre-Cambrian basement seem to be passively involved (Stose and Jonas 1935; Stose 1937).

POST-"APPALACHIAN" TRENDS

It is worthy of special attention that even the Triassic trough follows this trend. The southeast side of the trough is an overlap with few exceptions and dips gently outward. Detailed maps show this beautifully, especially the Honeybrook-Phoenixville quadrangles, Pa. (Bascom and Stose 1938), and the geologic map of Frederick County, Md. (Jonas and Stose 1939).

The outer side of the depression is bounded by many faults which effect a fairly consequent parallelism of that boundary with the trend of the Appalachian axes (Figs. 3 and 5). The general arc is thus followed by later structures but accompanied by dislocations which cut into the underlying pre-Triassic "basement." It is obvious that the "Appalachian" structure of the area has determined the trend of the trough which has been filled with Triassic sediments as well as its later deformation.

PRE-"APPALACHIAN" TRENDS OF THE BASEMENT

The author comprises under basement the following units: The Baltimore, Pickering, Pochuck, Byram and associated gneisses of Maryland,

Pennsylvania, and New Jersey with their intrusives. Included are all older crystallines which are unconformably overlain by basal Cambrian quartzites and sandstones like the Chickies quartzite and Hellam con-

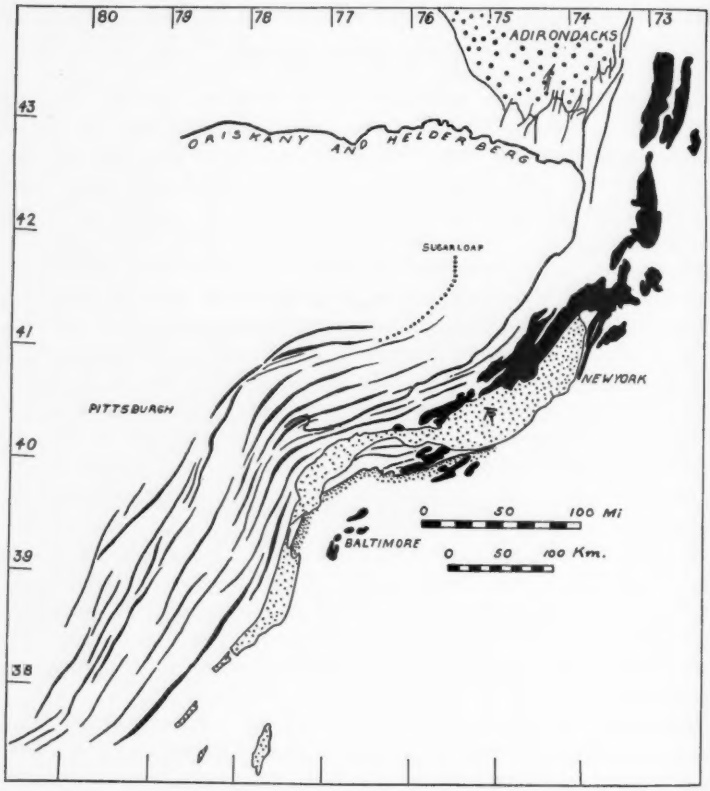

FIGURE 5.—*Tectonic map of the northern Appalachians*

Showing the arcuation between Baltimore and Pittsburgh and the continuation of the gneiss axis (solid black) from the northeast toward Baltimore. Heavy black lines are the fold axes. Stippled area nearest Baltimore is northwestern border of Glenarm series and trend of so-called "Martic overthrust end."

glomerate in Pennsylvania. The Wissahickon schist and other members of the Glenarm series are not included because that sequence rests on Baltimore gneiss and there is no locality known at which basal Cambrian strata can be observed unconformably on any member of the

Glenarm series. The only suggestion in that direction (Jonas 1924) has been revised in the publication of the geologic map of Frederick County (Jonas and Stose 1938).

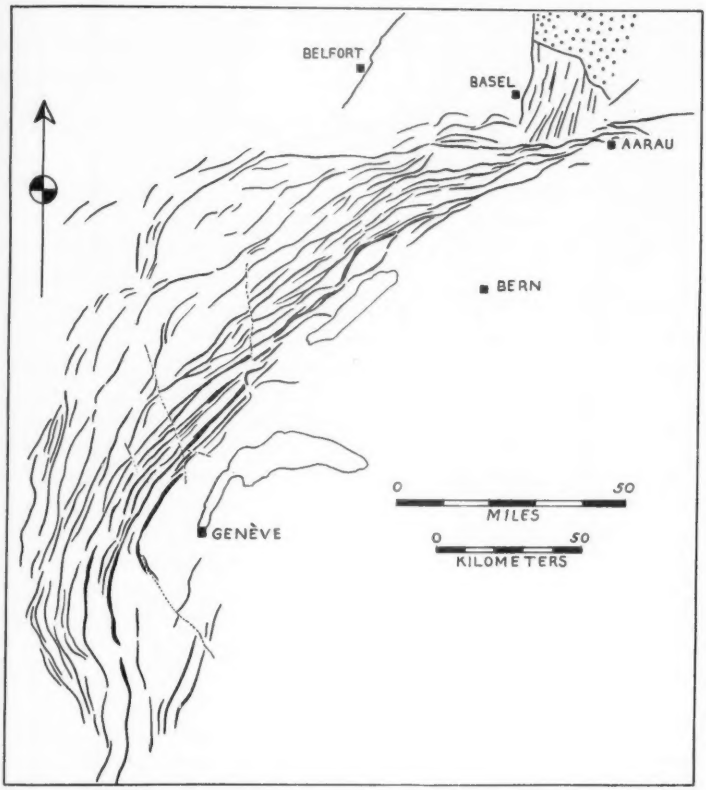

FIGURE 6.—*Tectonic map of the Jura mountains, Switzerland*
After Heim, for comparison with Figure 5.

The volcanic rocks and intrusives of South Mountain and Catoctin Mountain (Frederick County, Md.) are also excluded because they seem different from the Baltimore gneiss series. They will be discussed later (*See* Catoctin Volcanics).

The outline of this basement complex shows a general northeast-southwest trend. In Figure 5 the gneiss blocks are black. They are aligned

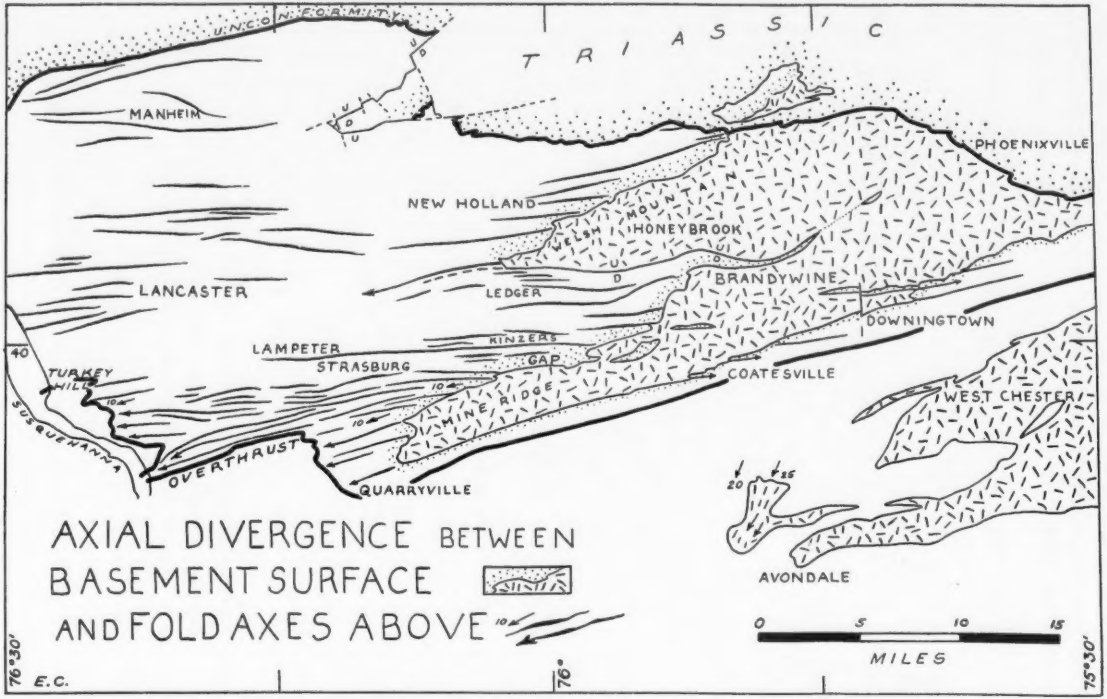

FIGURE 7.—*Axial divergence between basement surface and fold axes shown in Figure 5*
Figures show pitch of axes.

in a zone which does not participate in the outward swerve of Appalachian and Triassic structures but continues to Baltimore without deflection. Smaller units, as for instance the Reading uplift (Fig. 3), protrude westward a considerable distance from the axis (Fig. 9).

This alignment of gneiss units is highly significant in relation to the overlying Paleozoic structures and the amount of crustal shortening in each section.

DIVERGENCE BETWEEN PALEOZOIC AND PRE-CAMBRIAN AXES

PALEOZOIC FOLD AXES

On the general northeast-southwest trend of the basement units divergent directions of the later Appalachian arc are superposed at many localities. Depending on the locality within the arc the divergence may reach 90 degrees.

The most obvious examples occur in the Mine Ridge-Welsh Mountain uplift area (Bascom and Stose 1938). The northwestern slope of the two uplifts is the transgression surface of the basal Cambrian quartzites. The two slopes dip gently northwest and are roughly parallel. The southern boundary of the Welsh Mountain anticline is the steep Brandywine Manor fault (Stose 1937) which strikes east-west (Fig. 7). The southeastern boundary of the Mine Ridge anticline is a steep flexure with almost vertical dips.

The axes of the overlying Paleozoic folds trend east-west and thus form an angle with the strike of the northwestern slopes of the two uplifts. As a consequence the fold axes plunge down these slopes at gentle angles. The strike of the axes is constant and remains east-west even where the strike of the gneiss surface locally is north-south. East of Ledger and Kinzers (Fig. 7) the fold axes in the cover intersects the edge of the crystalline basement at an angle of 90 degrees.

West-southwest of Gap the Mine Ridge axis strikes southwest. From Gap due westward runs the Lampeter anticline. The distance between the two axes increases from one mile near Gap to 5 miles south of Lampeter, within a distance of 8 miles along the strike (Knopf and Jonas 1929b, Geol. map).

Structures of the Paleozoic cover have been preserved in depressed remnants of the younger rocks. North of Coatesville and Downingtown (Fig. 7) remnants of basal Cambrian quartzite strike east-west, transecting the general strike of the uplift-axis at an angle of 15 degrees. At Brandywine a depression in the gneiss surface is bounded by the plane of transgression of the basal Cambrian on the south side and the Brandywine Manor fault on the north. It seems as if the old crystallines had

Figures show pitch of axes.

failed in preserving their original trends and structures when new ones in diverging directions were superimposed. Where flexures and folds did not accomplish sufficient deformation, faults appeared.

The axial divergence which can be seen in the McCalls Ferry-Quarryville, Lancaster, New Holland, and Honeybrook quadrangles (Fig. 7) is also evident in the larger stratigraphic units.

The Triassic formations occur within the gneiss belt in New Jersey and eastern Pennsylvania. Between Philadelphia and Lancaster they "break through" the gneiss and swerve towards the north and west. In Lancaster county and from there on west the formations occur well outside and to the west—or in the "foreland"—of the gneiss belt.

In eastern Pennsylvania, the Welsh Mtn. and Mine Ridge uplifts are islands within Paleozoic sediments (Fig. 9). Near Baltimore the gneiss occurs in well defined uplifts which are surrounded by the crystalline schists of the Glenarm series. No well determined Paleozoic rocks occur anywhere near the domes at Baltimore.

CRYSTALLINE SCHISTS

The belt of crystalline schists is narrow in eastern Pennsylvania, and its northern boundary—the "Martic overthrust"—lies south of the Mine Ridge gneisses and well within the gneiss area. Near Quarryville this boundary leaps forward to the northwest and its trace in Figures 7 and 9 crosses the strike of the fold axes in an "overlap" of some 15 miles. From Turkey Hill at the Susquehanna River it follows the general trend of the fold axes to the west and southwest through York county (Stose and Jonas 1939) into Maryland (Figs. 7 and 9).

This forward surge of the schist boundary is either a structural or a stratigraphic overlap. The writer has mapped this critical area and a detailed report is in preparation (Cloos, Hietanen, and Guild, manuscript). A series of at least five thrusts, involving the schists and the underlying Paleozoic sandstones and limestones has been found at the western end of the Mine Ridge uplift. The largest horizontal transports occur between Quarryville and Turkey hill and die out rapidly toward the east.

The crystalline schists seem to have "broken through" into the foreland of the gneisses and across the Mine Ridge axis at the lowest point of the gneiss "barrier." There is no gneiss exposed between the Mine Ridge and the Phoenix dome north of Baltimore (Fig. 9). Between Quarryville and Baltimore many thousand feet of rocks of the Glenarm series are added to the column overlying the gneiss. It seems as if the low position of the gneissic basement had enabled the Glenarm series to "flow

through" this depression. Its depth is further emphasized by the presence of Ordovician Peach Bottom slate on both sides of the Susquehanna River. This is the only occurrence of such low metamorphic slate within the crystalline schist area. Near Peach Bottom and Delta the slates are in contact with garnet-kyanite schists with biotite porphyroblasts.

The writer is fully aware of the probability that the Wissahickon schist which now rests on Conestoga limestone between Turkey hill and Quarryville may have covered also a portion of the Mine Ridge area. It seems rather doubtful, however, that it should have also covered the area near Gap or farther northeast without leaving the slightest remnant behind. In the area between the Susquehanna River and the Mine Ridge, the author has found (report in preparation) a five-fold repetition of the sequence: Conestoga limestone, Vintage dolomite, Antietam schist. Proceeding from the river eastward the number of repetitions decreases to two north of Quarryville and the sequence becomes normal at Mine Ridge. In addition the petrographic investigation shows a decrease of metamorphism and the petrofabric investigation a decrease in the amount or rotational deformation in the same direction. The fivefold repetition is interpreted as due to thrusting since this sequence is well known elsewhere but never repeated.

The thrust surfaces are now intensely folded and their northern front trends east-west whereas Mine Ridge trends east-northeast. The northern thrust boundary, therefore, runs up the Mine Ridge anticline obliquely in the same way as the fold axes. Possibly only the southwestern end of the Mine Ridge area has ever been covered by the crystalline schists.

The forward surge of the Glenarm series west of Mine Ridge coincides with a series of additional folds and the general widening of the Appalachian folds in the area northwest of Harrisburg. This area shows the maximum amount of crustal shortening and bulging in the northern Appalachians.

TRENDS OF ISOGRADS

Isograds are surfaces indicating equal degrees of metamorphism. A detailed description and analysis of the isograds in Lancaster county is in preparation (Cloos, Hietanen, and Guild). The degree of metamorphism in the lower Paleozoic rocks, and crystalline schists is independent of the thickness of sediments above the gneiss basement and of the distance from the gneiss boundaries or any other stratigraphic boundaries; nor is it related to thrust surfaces or other tectonic boundaries.

Between New Holland and Honeybrook (Fig. 7), the basement is overlain by quartzites and fine-grained sandstones which are slightly re-

crystallized and carry small recrystallized biotite grains. Schistosity is very weak. From Kinzers to Lampeter an anticline of Antietam schist is covered by Vintage dolomite. The schist carries large biotites, and schistosity transects bedding. North of Quarryville and from there westward a coarse albite-muscovite schist appears with large biotite and albite porphyroblasts. The same formation which approaches the gneiss as a fine-grained sandstone west of Honeybrook appears as a coarse crystalline quartzose schist at the southwest end of Mine Ridge. This shows clearly that zones of equal metamorphism intersect the gneiss basement and its uplift axes. The isograds parallel the axes of the folds. The general increase of intensity of metamorphism has been mentioned by Knopf and Jonas (1929b, p. 128).

INTERNAL STRUCTURES OF THE PRE-CAMBRIAN GNEISS UNITS

The internal structures of the gneiss uplifts are little known. Broedel (1937) has described the domes near Baltimore, and recent work in New Jersey (Broughton 1940) has furnished some details. The Mine Ridge and Honeybrook uplifts have not been mapped in detail. The published maps (Bascom and Stose 1932, 1938; Knopf and Jonas 1929b) show neither strikes nor dips in the basement complex and the distribution of rocks as indicated in these maps is somewhat schematic and generalized and shows little of the internal structures. The general lack of good exposures for instance in the Mine Ridge uplift is distressing. Much careful work has been done in the Bryn Mawr area, but results have not been published.

The outlines of the gneiss domes near Baltimore seem to be influenced by their internal structures (Fig. 8B). Wherever primary structures (Balk 1937, p. 151; E. Cloos 1937, p. 49) form a dome the present outline is also dome-like. Accordingly, round and nearly circular and elongated domes have been observed (Broedel 1937, Pl. XXVII; Fig. 8B). Several of the domes consist of narrow cores of anticlines which are overturned southeastward (Cloos and Broedel 1940). The gneiss domes near Baltimore maintain strong individuality in contrast with the overlying Glenarm sediments. Fold axes in the latter are subparallel, and the uniformity of the general strike is only locally interrupted—either in the vicinity of the domes or near intrusive centers like the Baltimore gabbro (Cohen 1937) and the Port Deposit granite (Hershey 1937).

The gneiss domes are very strange units in an area which is otherwise determined by "Appalachian" trends (Fig. 8). They seem to be below these trends and the two directions coincide only in their immediate neighborhood.

LATER DEFORMATION OF THE BASEMENT UNITS

After the deposition of basal Cambrian sediments on the gneiss surface strong deformations altered the rocks above, the surface itself, and the gneiss below. Because of the intensity of the alteration it is now very difficult to separate the components which can be attributed to the individual phases of deformation. Large portions of the old floor on which the Paleozoic rocks were deposited have been more or less horizontal. Any deviation from that position is then due to later deformation. The degree of alteration within the gneiss complex is much more difficult to determine than that of its surface. The overlying Cambrian strata truncate the gneiss and its primary banding, petrographic boundaries, and many of its folds (Knopf-Jonas 1929a, p. 152).

Folding and faulting have deformed the basement complex. The Mine Ridge and Welsh Mountain anticlines are folds, partly accentuated by faults. The domes near Baltimore are largely folded and most likely the vertical uplifting component predominated. In the Woodstock anticlinorium (Fig. 8B), Howard County, Maryland, an anticline of Baltimore gneiss has been folded, overturned to the east, and pushed upward along reverse faults (Cloos and Broedel 1940).

G. W. Stose (1937) believes that the basement in the Welsh Mountain anticline has been thrust northward over the Paleozoic rocks east of New Holland and also that pre-Cambrian crystallines rest on later rocks near Reading, Pennsylvania (Stone and Jonas 1935). B. L. Miller and D. M. Frazer (1938) disagree with this view and present a more conservative interpretation. The author has mapped the area south of Lancaster, Pa., where five thrusts occur (Cloos 1940), and Howard county, Md. (1940) where the Baltimore gneiss has been pushed eastward over later rocks along reverse faults. However, the presence of far-travelled thrust sheets in which the pre-Cambrian gneisses are involved seems doubtful.

Normal faults have transected cover and basement at many localities and thus modified considerably the configuration of the basement surface. Several of the uplifts are partly, others wholly, bounded by faults. Primary deposition contacts and unconformities are in the minority. The intensity of later deformation of gneiss blocks is well illustrated by Balk (1936) in his description of the Clove quadrangle, N. Y., and by Broedel in the Baltimore area (1937, Pl. XXXII).

There is no doubt that portions of the basement complex have participated in later orogenies. They occur now in anticlinal cores and as faults blocks. How far they have been drawn into the deformation and to what extent their earlier structures have been obliterated cannot be determined without further detailed study. To an appreciable degree

primary structures have been preserved and they are disconformable with the later Appalachian trends.

CATOCTIN VOLCANICS

The South Mountain—Catoctin Mountain uplift in Maryland (Figs. 8 and 9) and its northward extension into Pennsylvania consists of a core of altered volcanic rocks, flanked and in part covered by the basal Cambrian Loudon and Weaverton formations. In Adams County (Stose 1932) and York County (Stose and Jonas, 1939), in Pennsylvania, this uplift is principally north and west of the Triassic belt. The only other occurrence of similar volcanic rocks is in the Hellam hills north of York Valley, Pa. (Stose and Jonas 1933). The Catoctin rocks continue southward across the Potomac River into Virginia. In the southern portion of the uplift the volcanic rocks are intruded by granitic rocks. Stose and Jonas (1939) are of the opinion that the volcanics rest on the granitic rocks unconformably. It is probable, however, that there are several generations of both rock types. The greenstones and purple slates of the Catoctin uplift are mostly highly schistose with enormous quantities of extremely well aligned stretched blebs and inclusions of all kinds. The linear stretching lies entirely within the cleavage planes and remains independent of the bedding. The direction of the schistosity and the stretching are remarkably constant over the whole area. Both structures extend far into the Paleozoic shales, sandstones, and limestones without change of direction, transecting their bedding at all angles. The boundary between the volcanics and the basal Cambrian is structurally meaningless and is transgressed by these structures (Fig. 8A). A detailed investigation of this area by L. B. Mather is in progress and a report is in preparation.

The greenstones, volcanics, granitic intrusions, and the Lower Cambrian rocks are traversed by "Appalachian" trends that surround them. They have become an important part of that system and differ only in their stratigraphic position and petrographic composition.

The basal Cambrian quartzites rest on the volcanics and locally a basal conglomerate is exposed containing pebbles of volcanic rocks (Bascom 1896). The Catoctin series is older than the basal Loudon and Weaverton formations, or the Chickies and Hellam in York County, Pennsylvania. Twenty miles east of the Catoctin uplift volcanic tuffs and greenstones of the same composition and with the same internal structures are interbedded with crystalline schists of the Glenarm series, quartzites, phyllites, and limestones. In this area volcanic slates are in contact with garnetiferous mica schists which contain biotite porphyroblasts.

COMPARISON OF CATOCTIN VOLCANICS AND BALTIMORE GNEISS

The author believes that the Catoctin volcanics do not belong to a general pre-Cambrian basement complex but have been deposited at a higher level above the gneiss series or may even belong to the lower Paleozoic complex. Baltimore gneiss does not occur in the Catoctin uplift axis.

There is a fundamental difference between the gneiss domes near Baltimore and the Catoctin uplift. Figure 8 illustrates the structures within and outside the uplifts. The data are taken from unpublished material of E. Cloos and L. B. Mather in the Catoctin area, and the geologic map of Howard County, Md. (Cloos and Broedel, 1940). The boundary of the pre-Cambrian cores in Figure 8 A and B are heavy black lines and the core itself is stippled.

The Catoctin volcanics are transgressed by the post-Cambrian structures of the whole region, whereas gneiss domes and anticlines clearly form the center of cleavage domes also in the adjacent schists. In Figure 8A the general Appalachian trends predominate but in Figure 8B local uplifts govern all structures. These local structures conform with the general Appalachian axes in a manner resembling a feldspar auge in an augen gneiss. Its internal structures are largely preserved—even if disturbed—but its shape is partly oriented and the foliae of the gneiss swing around it. Mantle structures flow around the domes but a short distance away Appalachian axial trends dominate again.

The petrographic difference is obvious: Metavolcanics with granitic intrusions on the one hand, and a mixed injection gneiss, well banded, including amphibolites and extensively intruded by granite, on the other.

The author believes that structurally and petrographically the Baltimore gneiss is a part of the same basement complex which underlies the uplifted areas in Pennsylvania and New Jersey. The Catoctin volcanics differ widely and in all probability they are above the Baltimore gneiss, and if the contact were exposed this surface would most likely be an important unconformity.

If Jonas' and Stose's (1939) contention is correct the granitic intrusions in the southern portion of the Catoctin uplift are older than the volcanics and possibly older than a portion of the southern basement complex, comparable to the northeastern gneisses.

The granitic complex has participated in the deformation of the Catoctin uplift, and its cleavage and intense linear stretching are parallel to that of the volcanics. The numerous dike-like bodies also parallel the cleavage. Thin sheets of volcanic rocks occur in the granite and thin layers of granite occur within the volcanics. The possibility that the

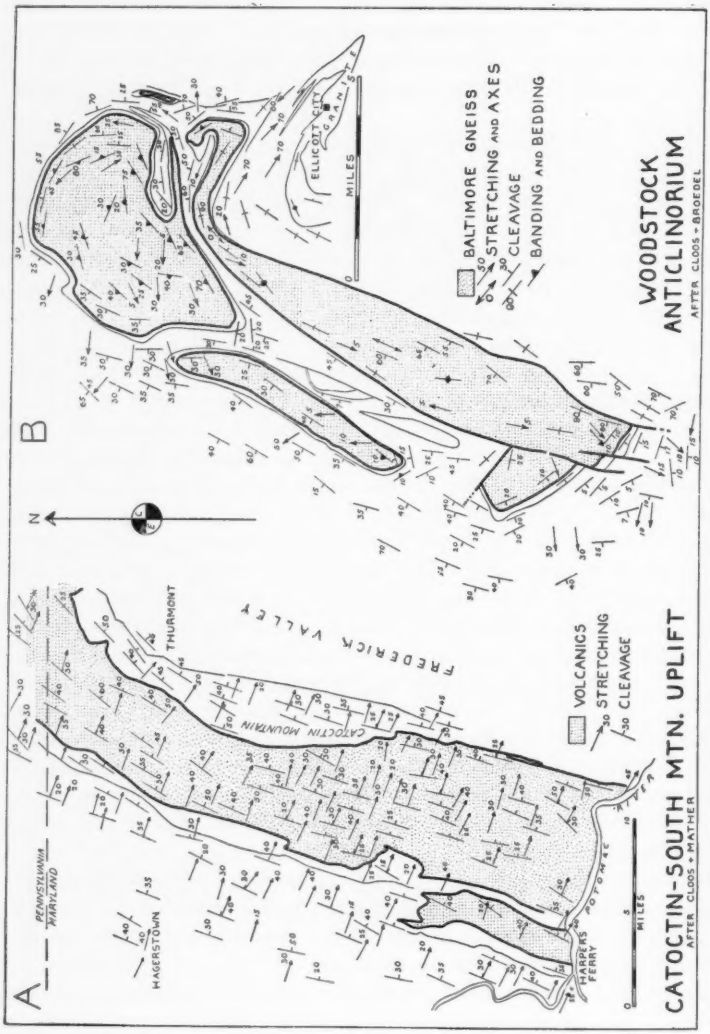

FIGURE 8.—*Structures in the Catoctin-South Mountain uplift, and Woodstock anticlinorium*

granitic rocks are not intrusive into the volcanics has never occurred to the author in the field and the problem seems to warrant detailed investigation.

SUMMARY AND CONCLUSIONS

The author has attempted a calculation of minimum crustal shortening ratios in a small portion of the northern Appalachians. The values are of the same order of magnitude as previous calculations by other workers. A fairly thorough search of the literature indicated that there were few areas for which similar calculations could be made because thorough work in that area had already been accomplished by the Second Geological Survey of Pennsylvania.

Only the horizontal component has been considered; thicknesses and wedges have not been calculated. The thickness may be found if it is possible to analyze the behavior of the basement and its role in Appalachian folding. If it participates extensively in large scale Appalachian folding the depth reached and the depth of the geosyncline remains a speculative matter. If folding remains largely above the basement the depth of folding is approximately equal to the depth of the geosyncline. Unfortunately neither value is known. The pre-Cambrian complex is still much discussed (Miller 1935; Mackin 1935; Cloos, Hietanen and Guild, manuscript) and it is not certain whether the Baltimore gneiss complex is the basement, or whether the Glenarm series and metavolcanic rocks have to be counted as the geosynclinal "floor", or a part of the folded geosyncline.

In this paper an attempt has been made to contribute some facts and thoughts to the discussion. The author realizes that the available data are not yet sufficient but feels justified in calling attention to some tectonic relations which should be studied.

Folding in "Appalachian" trends seems to take place selectively above the Baltimore gneiss complex, thus indicating a major boundary between the basement and its cover. The gneisses have been folded later but this deformation seems to be rather superficial since the more deeply eroded portions like the Honeybrook area show little or no remains of Appalachian trends and the less deeply eroded Baltimore gneiss domes show overturning and folding. At many localities an attempt to force the basement into later trends is only crudely accomplished by faulting.

The Glenarm series and the Catoctin volcanics participate in the Appalachian arc and should therefore be placed structurally above the gneiss complex.

Above the gneiss basement all formations and the trends of their metamorphism (isograds) surge forward through the "gneiss barrier" in

FIGURE 8.—*Structures in the Catoctin-South Mountain uplift, and Woodstock anticlinorium*

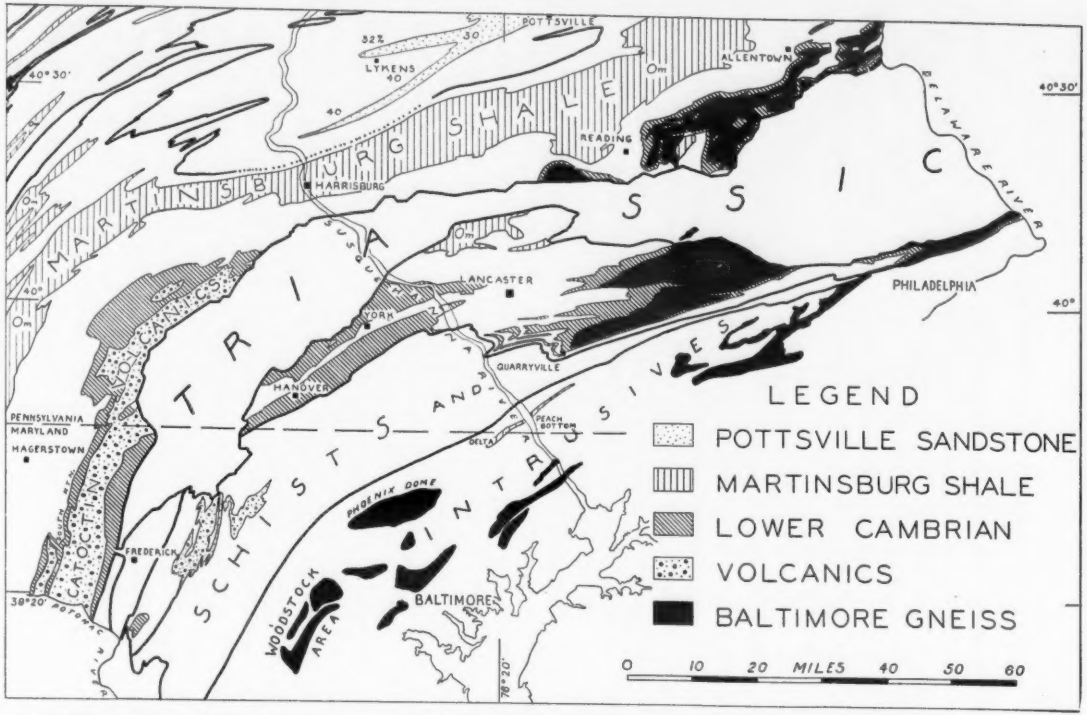

FIGURE 9.—*Generalized geologic map showing some important formations in eastern Pennsylvania and Maryland*
Key beds are emphasized by patterns and all others left white.

Pennsylvania from "behind" or within the gneiss to a position far ahead of the gneiss axis (Fig. 9). The Glenarm series and the lower Paleozoic rocks are thrust forward in southern Lancaster County, and their fold axes transgress the axes of the uplifts. Even the Triassic belt crosses the gneiss front and follows older pre-Triassic trends.

In contrast to the general northwest "surge" the gneisses are overturned locally and faulted intensely southeastward in Howard and Baltimore Counties, Md. The Mine Ridge uplift is an asymmetric dome with gentle northward dips along its northern flank and with steep (90° or overturned) dips along the southern margin. The Welsh Mountain anticline is also asymmetric with a gentle northern dip along its northwest flank and a steep or vertical fault along its southern side (Brandywine Manor fault). The gneiss complex itself does not indicate much northward push but shows rather tendencies toward southern or southeastward overturn. Stose assumes larger northward movements in Welsh Mountain (1937) and is inclined to see a considerable amount of passive participation of the basement in the Appalachian trends. (*See* also Stose and Jonas 1935; Miller and Frazer 1936).

The author thinks that a systematic comparison of structures, their directions, development and age relations may be an aid in deciphering the geologic history of the intricate area where later deformations and metamorphism have obliterated much organic fossil evidence and left only structures and minerals in their place. The area is much discussed and it seems that there is need for much more detailed work.

WORKS TO WHICH REFERENCE IS MADE

Ashburner, C. A. (1886) *Second report of progress in the anthracite coal regions,* Pa. Geol. Survey, Ann. Rpt.

Ashley, G. H. (1931) *A syllabus of Pennsylvania geology and mineral resources,* Pa. Geol. Survey Bull. G. 1.

Balk, Robert (1936) *Structural and petrologic studies in Dutchess County, New York,* pt. I. *Geologic structure of sedimentary rocks,* Geol. Soc. Am., Bull., vol. 47, p. 685-774.

——— (1937) *Structural behavior of igneous rocks,* Geol. Soc. Am., Memoir 5.

Bascom, Florence (1896) *The ancient volcanic rocks of South Mountain, Pennsylvania,* U. S. Geol. Survey, Bull. 136.

———, and Stose, G. W. (1932) *Coatesville-West Chester folio,* U. S. Geol. Survey Folio 223.

——— — ——— (1938) *Geology and mineral resources of the Honeybrook and Phoenixville quadrangles, Pennsylvania,* U. S. Geol. Survey, Bull. 891.

Broedel, Carl H. (1937) *The structure of the gneiss domes near Baltimore, Maryland,* Md. Geol. Survey, vol. XIII, pt. III, p. 149-184.

Broughton, John G. (1940) *Comparison of Paleozoic and pre-Cambrian structures in northwestern New Jersey.* Dissertation, Johns Hopkins Univ.

Bucher, W. H. (1933) *The deformation of the earth's crust,* Princeton Univ. Press.

FIGURE 9.—Generalized geologic map showing some important formations in eastern Pennsylvania and Maryland
Key beds are emphasized by patterns and all others left white.

Chamberlin, R. T. (1910) *The Appalachian folds of central Pennsylvania,* Jour. Geol., vol. 18, p. 228-250.

——— (1919) *The building of the Colorado Rockies,* Jour. Geol., vol. 27, pt. I, p. 145-164, pt. II, p. 225-251.

Cloos, Ernst (1937) *The application of recent structural methods in the interpretation of the crystalline rocks of Maryland,* Md. Geol. Survey, vol. XIII, pt. 1, p. 25-100.

———, and Broedel, C. H. (1940) *Geologic map of Howard County, Maryland,* Md. Geol. Survey.

———, Hietanen, Anna, Guild, P. *The geology of the Glenarm series and the Martic overthrust in Pennsylvania and Maryland,* Geol. Soc. Am., Special Papers series, (manuscript).

Cohen, Charles J. (1937) *Structure of the metamorphosed gabbro complex at Baltimore,* Md. Geol. Survey, vol. XIII, p. 215-235.

Frazer, D. M. (1938) *Contributions to the geology of the Reading Hills,* Pa., Geol. Soc. Am., Bull., vol. 49, p. 1199-1212.

Heim, Albert (1919) *Geologie der Schweiz,* 3 vols. Leipzig, Chr. H. Tauchnitz.

Hershey, H. G. (1937) *Structure and age of the Port Deposit granodiorite complex,* Md. Geol. Survey, vol. XIII, p. 109-146.

Jonas, A. I. (1924) *Pre-Cambrian rocks of the western piedmont of Maryland,* Geol. Soc. Am., Bull., vol. 35, p. 355-364.

———, and Stose, G. W. (1926) *"New Holland quadrangle,"* Top. and Geol. Atlas of Pa., No. 178.

——— — ——— (1930) *"Lancaster quadrangle,"* Top. and Geol. Atlas of Pa., No. 168.

——— — ——— (1938) *Geologic map of Frederick County, Maryland,* Md. Geol. Survey.

——— — ——— (1939) *Age relation of the pre-Cambrian rocks in the Catoctin Mountain-Blue Ridge and Mount Rogers anticlinoria in Virginia,* Am. Jour. Sci., vol. 237, p. 575-593.

Keith, Arthur (1894) *Geology of the Catoctin belt,* U. S. Geol. Survey, 14th Ann. rpt., pt. 2, p. 285-395.

——— (1923) *Outlines of Appalachian structures,* Geol. Soc. Am., Bull., vol. 34, p. 309-380.

——— (1928) *Structural symmetry in North America,* Geol. Soc. Am., Bull., vol. 39, p. 321-386.

King, P. B. (1937) *Geology of the Marathon region, Texas,* U. S. Geol. Survey, Prof. Paper 187.

Knopf, E. B., and Jonas, A. I. (1929a) *The geology of the crystalline rocks of Baltimore County,* Md. Geol. Survey, Baltimore Cty. report, p. 97-216.

——— — ——— (1929b) *Geology of the McCalls Ferry-Quarryville district, Pennsylvania,* U. S. Geol. Survey, Bull. 799.

Mackin, J. H. (1935) *The problem of the Martic overthrust and the age of the Glenarm series in southeastern Pennsylvania,* Jour. Geol., vol. 43, no. 4, p. 356-380.

Mansfield, G. R. (1927) *Geography, geology, and mineral resources of part of southeastern Idaho,* U. S. Geol. Survey, Prof. Paper 152.

Marshall, John (1937) *The structure and age of the volcanic complex of Cecil County, Maryland,* Md. Geol. Survey, vol. XIII, p. 189-211.

Mathews, E. B. (1904) *The structure of the piedmont plateau as shown in Maryland,* Am. Jour. Sci., vol. XVII, p. 141-159.

Miller, B. L. (1935) *Age of the schists of the South Valley Hills, Pennsylvania,* Geol. Soc. Am., Bull., vol. 46, p. 715-756.

——, and Fraser, D. M. (1936) Comment on *"Highlands near Reading, Pennsylvania; an erosion remnant of a great overthrust sheet,"* by George W. Stose and Anna I. Jonas, Geol. Soc. Am., Bull., vol. 46, p. 2031-2038.

Stose, George W. (1932) *Geology and mineral resources of Adams County, Pennsylvania,* Pa. Geol. Survey, Bull. C 1.

—— (1933) *Geologic map of Pennsylvania,* six inches equal one mile. Pa. Geol. Survey.

——, and Jonas, A. I. (1933) *Geology and mineral resources of the Middletown Quadrangle, Pennsylvania,* U. S. Geol. Survey, Bull. 840.

—— (1937) *Structure of the Honeybrook uplift, Pennsylvania,* Geol. Soc. Am., Bull., vol. 48, p. 977-1000.

—— — —— (1935) *Highlands near Reading, Pennsylvania; an erosion remnant of a great overthrust sheet,* Geol. Soc. Am., Bull., vol. 46, p. 757-779.

—— — —— (1939) *Geological map of York County, Pennsylvania,* Pa. Geol. Survey.

Van Hise, C. R. (1898) *Estimates and causes of crustal shortening,* Jour. Geol., vol. 6, p. 10-65.

White, I. C. (1881) *Geologic map of Susquehanna and Wayne counties,* etc., Second Geol. Survey, Grand Atlas, Div. 1, pt. 1, Sheet No. XLI.

Johns Hopkins University, Baltimore, Md.

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Northern Anthracite Fields

No. of Section*	Formation or bed used	Length of Section (Ft.)	Reduction Ratio: $\frac{b}{a}$	Reduction $100-\frac{b}{a}.100(\%)$	Remarks
K	Pottsville base	15,000	36.8/37.3	1	Northernmost section
J	Pottsville base	14,500	36.5/36.5	0	
	Bottom coal bed	5,500	13.7/13.5	0	
I	Pottsville base	19,200	48/48.5	1	
H	Pottsville base	27,500	68.7/69.5	1	
G	Mauch Chunk	30,000	76.5/77.3	1	
	Dunmore No. 3	24,500	61/62	1.3	
F	Mauch Chunk	21,500	53.8/54.8	1.2	
	Red ash bed	18,800	47.7/48.2	1	
E	Mauch Chunk	41,000	64.8/65.5	1	
	Red ash bed	17,000	43/43.3	0.5	
D	Red ash bed	20,100	52.3/52.8	0.7	
C	Pottsville base	32,000	81.4/83.0	2	
	Eleven foot bed	21,500	54.8/56	2.5	
B	Pottsville base	27,000	67.3/73.4	8	
	Baltimore bed	21,500	54.5/60	9	
A	Mauch Chunk top	17,500	44/48	8	K-A general sections from NE to SW
	Buck Mt. bed	15,000	37.6/41	8	
1	Pottsville base	3,100	7.8/8.8	11	
2	Pottsville base	4,000	10.1/11.4	11.5	
	Buck Mt. bed	3,800	9.2/10.4	11.5	
3	Pottsville base	8,000	20.2/22	8	
	Red ash bed	7,200	18.0/19.5	7.5	
4	Pottsville base	10,000	25.2/28.2	10.5	
	Buck Mt. bed	9,000	22.5/25.5	11.5	

Eastern Middle Field

General Sections

1	Top of Pottsville Conglomerate	3,900	9.8/10.7	8.2	
2	Top of Pottsville Conglomerate	4,300	10.8/11.8	8.5	
3	Top of Pottsville Conglomerate	5,600	14/17	17.8	
4	Top of Pottsville Conglomerate	4,400	11/12.7	13	
5	Top of Pottsville Conglomerate	3,100	7.7/8.3	7	

Special Sections

6	Buck Mt. bed	2,800	7/7.5	6.5	
7	Top of Pottsville Conglomerate	2,700	6.8/8.2	17	
8	Too short and exposure poor				
9	Mammoth bed	1,120	2.8/3.4	17.5	Short
10	Below Mammoth bed	2,000	5/6	16.5	
11	Mammoth bed	1,900	4.8/6	20	
12	Buck Mt. bed	2,200	5.5/6.6	16.5	
13	Too short				
14	Too short				
15	Buck Mt. bed	3,300	8.4/9.8	14.4	
16	Buck Mt. bed	3,750	9.4/10	6	
17	Buck Mt. bed	4,275	10.5/11.7	10.5	

Eastern

Sp

No. of Section*	Formation or bed used	Le Secti
18	Mammoth bed	
19	Mammoth bed	
20	Mammoth bed	
21	Buck Mt. bed	
22	Pottsville Conglomerate (top)	
23	Pottsville top Mammoth bed	
24	Pottsville top Mammoth bed	
25	Pottsville top Mammoth bed	
26	Incomplete	
27	Pottsville top Mammoth bed	
28	Mammoth bed	
29	Wharton bed Mammoth bed	
30	Buck Mt. bed	
31	Buck Mt. bed	
32	Pottsville base	
33	Pottsville base	
34	Buck Mt. bed	
35	Buck Mt. bed	
36	Buck Mt. bed	
37	Poor exposures	
38	Buck Mt. bed	
39	Buck Mt. bed Wharton bed	
40	Wharton bed Buck Mt. bed	
41	Wharton bed	
42	Wharton bed Buck Mt. bed	
43	Buck Mt. bed	
44	Buck Mt. bed	
45	Wharton bed	
46	Buck Mt. bed	
47	Too short	
48	Buck Mt. bed	
49	Buck Mt. bed Buck Mt. bed	
50	Too short	

West

1	Buck Mt. bed	
2	Buck Mt. bed	
3	Buck Mt. bed	
4	Buck Mt. bed	
5	Mammoth bed Mammoth bed	
6	Mammoth bed Mammoth bed	

* Numbers are the same as in the Anthracite Reports of the Second Geol. Survey of Pa.

TABLE 1.—*Cross sections through the anthracite field of eastern Pennsylvania*

Based on the reports of the Second Geological Survey

Eastern Middle Field (continued)

Special Sections (continued)

No. of Section*	Formation or bed used	Length of Section (Ft.)	Reduction Ratio: $\frac{b}{a}$	Reduction $100-\frac{b}{a}.100(\%)$	Remarks
18	Mammoth bed	2,000	5/5.5	9	
19	Mammoth bed	1,920	4.8/5.5	12.5	
20	Mammoth bed	1,800	4.5/5.2	13.5	
21	Buck Mt. bed	3,400	8.5/9.5	10.2	
22	Pottsville Conglomerate (top)	4,000	10/11	10	
23	Pottsville top	3,500	8.8/10	12	
	Mammoth bed	2,000	5/6.2	19	
24	Pottsville top	4,000	10/13	23	
	Mammoth bed	2,000	5/7.2	30	
25	Pottsville top	4,600	11.5/14.5	20.5	
	Mammoth bed	2,500	6.2/8.8	29.5	
26	Incomplete				
27	Pottsville top	6,500	16.8/18.6	9.5	
	Mammoth bed	3,600	9.03/10.5	14	
28	Mammoth bed	3,000	7.5/8.3	9.5	
29	Wharton bed	5,600	14/14.6	3	
	Mammoth bed	2,000	5/5.3	6	
30	Buck Mt. bed	2,700	6.8/8	15	
31	Buck Mt. bed	1,100	2.8/3	6.5	Very short section
32	Pottsville base	1,750	4.4/5	12	Very short section
33	Pottsville base	2,800	7/8.4	16.5	
34	Buck Mt. bed	2,400	6/7	14.5	
35	Buck Mt. bed	3,800	9.6/11.2	14	Thrust 0.5%
36	Buck Mt. bed	1,700	4.3/5.7	24.5	
37	Poor exposures				
38	Buck Mt. bed	9,000	22.5/25.3	15	
39	Buck Mt. bed	6,000	14.8/16.8	12	
	Wharton bed	4,000	10/12	16.5	
40	Wharton bed	4,400	11/15.2	27	Some reverse faulting
	Buck Mt. bed	7,000	17.5/22.2	21	
41	Wharton bed	3,800	9.5/15.25	37.8	
42	Wharton bed	1,600	4.2/5.5	23.5	
	Buck Mt. bed	2,400	6.2/8.2	24.5	
43	Buck Mt. bed	2,000	5/7	28.5(?)	Reverse faulting
44	Buck Mt. bed	1,600	4/6.8	41(?)	Tight folds, not completely exposed
45	Wharton bed	2,800	7.3/7.8	6.5	
46	Buck Mt. bed	1,400	3.5/5	30	
47	Too short				
48	Buck Mt. bed	1,840	4.6/6.2	26	
49	Buck Mt. bed	1,200	3/6.2	52(?)	Overturned south
	Buck Mt. bed	1,300	2/3.2	37(?)	
50	Too short				

Western Middle Anthracite Field

1	Buck Mt. bed	8,000	22/24	8.5	Very good section along tunnel
2	Buck Mt. bed	9,000	22.2/28.3	22	Excellent
3	Buck Mt. bed	9,000	22.8/27.5	17	Good section
4	Buck Mt. bed	10,000	25.8/34.8	25.5	Good section
5	Mammoth bed	2,000	5/8.2	39	
	Mammoth bed	4,000	10/15.7	36.5	Portion of above section
6	Mammoth bed	1,400	3.5/6.4	45.5	
	Mammoth bed	5,500	13.8/16.3	15	Portion of above section

Western Middle Anthracite Field (continued)

No. of Section*	Formation or bed used	Length of Section (Ft.)	Reduction Ratio: $\frac{b}{a}$	Reduction $100-\frac{b}{a}.100(\%)$	Remarks
7	Too short and incomplete				
8	Mammoth bed	800	2/5	60	One sharp fold exposed
	Projected Mammoth bed	9,600	24/36.6	34	
9	Mammoth bed	8,800	22.3/33.5	33	Shortened by thrust
10	Buck Mt. bed	1,300	3.3/6.4	48	One syncline only
11	Mammoth bed	3,380	8.4/14.5	42	One syncline
	Mammoth bed	4,200	9.6/12.5	31	Average of total section
12	Mammoth bed	9,000	22.2/26.8	17	
	Mammoth bed	4,400	11.2/16.8	33.5	Overturned and faulting
	Mammoth bed	4,500	8.2/12.5	34	
13	Buck Mt. bed	22,000	55/65.3	16	Average projected
14	Buck Mt. bed	16,000	39.8/46	14	
15	Mammoth bed	13,200	33.7/42.2	20	Partly projected
16	Mammoth bed	11,600	29.3/36	18	
17	Mammoth bed	800	20.2/25.7	22	Very short
18	Mammoth bed	600	12.7/15.5	18	Very short

Southern Anthracite Field (East)

1	Buck Mt. bed	2,500	3.5/7	50	Double fold, tunnel
2	Mammoth bed	4,000	7.2/14	50	Tunnel exposure
3	Mammoth bed	3,400	5.3/8	34	
	5 ft. coal bed		4/9	56	
4	Mammoth bed	4,000	10/16	37	
5	Mammoth bed	5,000	13/21	38	Mostly interpreted
6	Mammoth bed	5,000	12.7/19.2	34	
	G-bed	3,600	9/15	40	
7	Mammoth bed	9,000	19/27.5	31	Mostly projected
8	Mammoth bed	8,000	18.5/29	36	
	F-bed		2.8/7	60	Sharp fold
	G-bed		1.2/2.8	57	
9	Mammoth bed	7,500	18.5/29.5	37	Generalized
	Mammoth bed		4.4/6	27	Double fold
10	Mammoth bed	7,500	17.8/28	36	Mostly projection
11	Too little information				
12	Mammoth bed	5,500	13.5/22.7	40	Much projection
13	Buck Mt. bed	6,000	14.6/21.4	32	Much projection
14	Primrose bed	7,000	13.5/20.2	33	
15	Buck Mt. bed	11,000	26.5/34	22	
16	Buck Mt. bed	13,000	32/44	27.5	
			2.8/3.7	24	
17	Mammoth bed	23,000	40.4/53.8	25	
	Diamond bed		23.1/31.5	26.5	
18	Mammoth bed	22–23,000	53.5/72	26	
	Peach Mt. bed		4/6	34	
	Mammoth bed		6/10	40	
19	Buck Mt. bed	22,000	11.6/18.2	36	
	Buck Mt. bed		48.2/64.5	25	Generalized
20	Buck Mt. bed	30,000	20/26.6	25	
	Buck Mt. bed		34.2/5.1	33	
	Daniel bed		12.2/17	28	
21	Buck Mt. bed	30–35,000	47.5/67	29	Generalized
22	Buck Mt. bed	30,000	34.2/51	33	Generalized
23	Buck Mt. bed	18,000	32.5/50.2	35	
	Lykens Valley—5		2/3.5	42.5	North end of 23
	Lykens Valley—5		4/7.5	47	

Southern Anthracite Field (East—continued)

No. of Section*	Formation or bed used	Length of Section (Ft.)	Reduction Ratio: $\frac{b}{a}$	Reduction $100-\frac{b}{a}.100(\%)$	Remarks
24	Buck Mt. bed		41.5/59.5	30	Generalized
	Buck Mt. bed		14.2/19.3	26	
25	Buck Mt. bed	6,560	16.5/23	28	
26	Lykens Valley—5	8,800	22.2/30	26	
27	Little bed	7,400	18.5/24.8	25	
	Mammoth bed		6.8/11	28	Good average
	Primrose bed	2,300	5.8/9.7	40	
28	Lykens Valley	5,000	12.8/18.5	31	Good, tunnel exp.
29	Pottsville base	5,800	14.5/25	42	Generalized
	Heister bed	2,200	5.6/11	49	
30	Heister bed	1,800	4.5/8.6	48	
	Pottsville base	6,000	15/26.5	43	
31	Pottsville base	4,800	12/19.5	38	

Eastern Middle Anthracite Field

General Sections

1	Top of Pottsville	3,920	9.8/10.7	8	
2	Top of Pottsville	4,320	10.8/11.8	8	
3	Top of Pottsville	5,600	14/17	17.8	
4	Top of Pottsville	4,400	11/12.7	13	
5	Top of Pottsville	3,100	7.7/8.3	7	

Special Sections

6	Buck Mt. bed	2,800	7/7.5	6.5	
7	Top of Pottsville conglomerate	2,720	6.8/8.2	17	
8	Not complete				
9	Mammoth bed	1,120	2.8/3.4	18	
10	Below Mammoth bed	2,000	5/6	16.5	
11	Mammoth bed	1,920	4.8/6	20	
12	Buck Mt. bed	2,200	5.5/6.6	16.2	
13	Incomplete				
14	Incomplete				
15	Buck Mt. bed	3,360	8.4/9.8	14	
16	Buck Mt. bed	3,760	9.4/10	6	
17	Buck Mt. bed	4,200	10.5/11.7	10	
18	Mammoth bed	2,000	5/5.5	9	
19	Mammoth bed	1,920	4.8/5.5	12.5	
20	Mammoth bed	1,800	4.5/5.2	13.5	
21	Buck Mt. bed	3,400	8.5/9.0	6	
22	Pottsville conglomerate	4,000	10/11	9	
23	Pottsville conglomerate (top)	3,400	8.8/10	12	
	Mammoth bed	2,000	5/6.2	19	
24	Top of Pottsville conglomerate	4,000	10/13	23	
	Mammoth bed	2,000	5/7.2	30	
25	Top of Pottsville conglomerate	4,400	11.5/14.5	20.5	
	Mammoth bed	2,480	6.2/8.8	29.5	
26	Not passable				
27	Top of Pottsville	6,720	16.8/18.6	9.5	
	Mammoth bed	3,600	9.03/10.5	14	
28	Mammoth bed	3,000	7.5/8.4	10.5	
29	Wharton bed	5,600	14/14.6	4	
	Mammoth bed	2,000	5/5.3	6	

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STATUS OF GEOLOGICAL AND GEOPHYSICAL INVESTIGATIONS ON THE ATLANTIC AND GULF COASTAL PLAIN

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ABSTRACT

Major features of the structure of the Gulf and Atlantic Coastal Plain of the United States are summarized. They include a postulated geosyncline of proportions comparable to those of the Appalachian of Paleozoic time. The attitude of the basement floor in Arkansas and along the Atlantic coast as inferred from wells which penetrated it and geophysical data on basement configuration are discussed, with emphasis on seismic methods. Seismic velocities in the basement rocks and younger sediments, possible depth limitations of the seismic method, and tests of the method are described. Seismic investigations on the Atlantic Coastal Plain are reviewed, with special reference to work at sea near the edge of the continental shelf, where attempts to map the basement rocks gave inconclusive results.

GENERAL HISTORY

The structure of the Gulf Coastal Plain has been determined in considerable detail at many points where oil and gas have been produced or sought. Surface geologic mapping, subsurface geologic mapping from drill records, and geophysical methods have contributed information. A few wells have supplied data on the Atlantic Coastal Plain in less detail. These have been supplemented by dredging along the walls of submarine canyons in the continental shelf (Stetson, 1936).

In broad outline, the early histories of the Gulf and Atlantic Coastal Plains appear to have many similarities. At the beginning of the Cretaceous the regions were land areas having the low rolling topography which characterizes a peneplain. Remnants of this surface are found today in the Piedmont which separates the Atlantic Coastal Plain from the folded Appalachians. Present relief of the exposed Piedmont does not exceed a few hundred feet and the absolute elevations range from sea level at New York City to 2,000 ft. in northwestern Georgia. Rocks of the Piedmont include pre-Cambrian, Cambrian, Ordovician, late Paleozoic, and Triassic, the last in isolated basins. The pre-Cambrian consists largely of granites and metamorphic rocks of complex structure (Billings and Williams, 1932). The Cambrian and Ordovician rocks are folded sediments in places metamorphosed by heat and pressure and, in Georgia and the Carolinas, intruded by large bodies of granite during the Appalachian Revolution. Lava flows and red sandstones and shale dominate the Triassic formations.

A Lower Cretaceous sea encroached on this basement complex from the southwest (Stephenson, 1928) and ultimately covered most of Texas and Louisiana, southwestern Arkansas, and possibly southwestern Mississippi. After an interval of erosion, a major inundation contemporaneous with that of the Rocky Mountain geosyncline covered the entire Gulf and most of the Atlantic Coastal Plain as far northeastward as Georges Bank. Downwarping of the Mississippi embayment allowed the sea to spread northward to the southern part of Illinois. There followed a series of marine transgressions and regressions which distributed the Coastal Plain sediments as they are found today. The present inner margin of those sediments is shown on the map of Figure 1. An approximation of the outer margin is generally believed to be represented by the break in slope at the continental margin, roughly defined by the 100-fathom line on the map.

STRUCTURE

The Coastal Plain sediments are characterized by low-angle coastward dips which are modified by minor regional folds and, at least in the

Gulf section where information is most detailed, by extensive faulting in certain areas.

PRINCIPAL AXES OF FOLDING

Principal axes of anticlinal and synclinal warping are shown in Figure 1, following Stephenson. Anticlines are indicated by dip arrows pointing

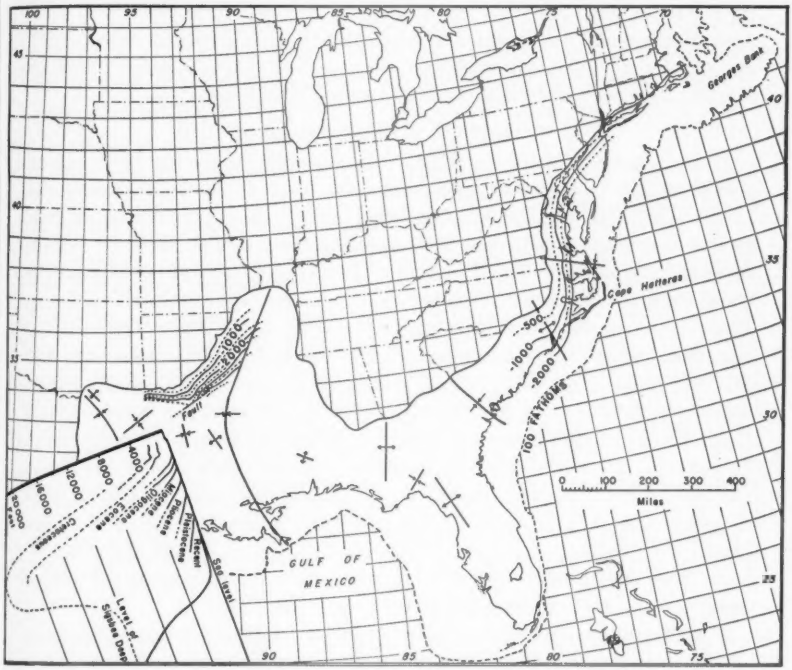

FIGURE 1.—*Map of Atlantic and Gulf Coastal Plain*

Inner margin of Coastal Plain sediments shown by heavy line. Structural trends according to Stephenson; cross section of Gulf Coast geosyncline according to Barton, Ritz, and Hickey; basement contours in Arkansas according to Spooner. Anticlines are indicated by dip arrows pointing away from axis lines and synclines by the reverse.

away from axis lines and synclines by the reverse. The Mississippi embayment is a broad sediment-filled trough warped downward slightly and plunging gently toward the Gulf of Mexico. Florida is warped upward along an axis trending north-northwest. The Chattahoochee River flows southward along the Georgia-Alabama line on the axis of a broad upwarp. A shallow syncline separates the Chattahoochee and Florida upwarps, plunging southwestward at a low angle. The Hatchetigbee anti-

cline trends northwest-southeast in the west-central part of Alabama between the Chattahoochee upwarp and the Mississippi downwarp. It is 50 miles long by 20 miles wide, and beds involved show elevations of 600 to 700 ft. above their normal positions. Features not shown in Figure 1 include a fault in Wilcox County, Alabama, on which a displacement of 300 ft. has been reported, and numerous minor faults in Upper Cretaceous rocks in the same county. An upwarp of the anticlinal nose type with axis trending 32° southwest has been described at Jackson, Mississippi, and southward dipping monoclines are reported near Vicksburg, Warren County, and near Eldorado, Yazoo County, Mississippi.

FAULTING IN GULF COASTAL PLAIN

Major features west of the Mississippi embayment are the Monroe and Sabine uplifts of northern Louisiana, the East Texas embayment, the Llano uplift, and the Rio Grande embayment. The Balcones fault zone bounds the inner edge of the Coastal Plain from Uvalde County, Texas, eastward and northeastward to Austin, Travis County, and extends northward from there into the Coastal Plain at least as far as Bell County. The faults are normal with downthrow to the south and southeast and combined vertical displacements probably amount to a maximum of 1,000 ft. The Mexia-Powell fault zone is traceable from Uvalde County in the southwest to Titus County, Texas, in the northeast. As far as Travis County it parallels the Balcones fault zone at a distance of 10 to 15 miles. In general, it coincides approximately with the outcrop of the basal Eocene beds. The faults range up to 25 miles or more in length, with displacements of as much as 600 ft. Graben faults associated with the Cretaceous-Tertiary contact, with displacements of as much as 400 ft., have been traced across part of southern Arkansas and are considered genetic associates of the Mexia-Powell zone in Texas.

Faulting in Coastal Plain sediments is generally attributed to mobility of the basement complex, since the sediments represent a relatively thin veneer which would react readily to movement of the foundation. Evidence of present activity is seen in major earthquakes at New Madrid, Missouri, in 1811 and Charleston, South Carolina, in 1886, as well as intermittent but continuing minor disturbances throughout the area, with some of the recent ones in the vicinity of New York City.

CENTRAL ATLANTIC COASTAL PLAIN

Darton (1938) has investigated structural relations and the history of Cretaceous and Tertiary formations of the Central Atlantic Coastal Plain in eastern Maryland. He reports,

"Although in general the strata thicken and also dip to the southeast, there are many local irregularities, most of them resulting in irregular truncation or bevelling of the deposits, and in places there is considerable channelling. The deformations have

been mostly widespread tilting, but the rate, direction, and extent of uplift vary in different parts of the region, as well as in different formations, and there is some slight flexing. . . . It is evident that in middle Miocene time the entire area now occupied by Washington and Baltimore was submerged by the sea."

GULF COAST GEOSYNCLINE

Barton, Ritz, and Hickey (1933) have assembled evidence that in the Gulf Coast area of Texas and Louisiana there is a geosyncline of proportions comparable to those of the Appalachian of Paleozoic time. A cross section of the structure is inset in the map of Figure 1, along a line from Shreveport, Louisiana, north of the Sabine uplift, through Jennings, Louisiana, to the Sigsbee Deep in the Gulf of Mexico. An abstract of the data, by those authors, states:

"The presence of a geosyncline along the Gulf Coast of Texas and Louisiana is indicated both by geologic and geophysical data. Formations which are exposed at the surface or in drilling are known in general to dip Gulfward. The known stratigraphic thickness of formations plus extrapolation below reach of the drill indicate that the depth to the basement at Houston is greater than 20,000 ft., at Jennings, La., greater than 25,000 ft., and south of New Orleans greater than 30,000 ft. Seismic prospecting indicates that the basement certainly is deeper than 15,000 ft. and, less definitely, that it is deeper than 20,000 ft. The torsion-balance data indicate that the base of the salt core of the salt domes lies at a depth of 17,000-20,000 ft. or more in the Houston district. The depth of the Gulf of Mexico in the Sigsbee Deep is 12,500 ft. and throughout most of its area is less than 10,000 ft. The thickness of the post-Lower Cretaceous sediments in the central part of the Gulf of Mexico presumably is not more than a very few thousand feet. The basement of the post-Lower Cretaceous beds in the Gulf Coast must be warped down at least 5,000 ft. at Houston and at least 15,000 ft. in the area south of New Orleans. A regional trough of gravity minimum lies axially along the Gulf Coast. Calculations suggest that it is best explained by a geosyncline on the basement plus a progressive character change of the basement from granitic under the land to basaltic under the Gulf of Mexico. A study of the depression of the basement to compensate sedimentation suggests that the trough line of the geosyncline should lie nearly at the present coast line; and that is approximately the position which the gravitational calculations indicated for the actual position of the trough line. Although isostatically negative, the area has been one of continued subsidence; and the rate of subsidence seems to have kept even pace with the rate of sedimentation."

BASEMENT FLOOR IN ARKANSAS

Contours on the basement flour underlying Coastal Plain sediments in Arkansas are shown in Figure 1, following Spooner (1935). They are based on the records of 85 wells which penetrated the basement complex at depths ranging down to 4,740 ft. below sea level. From the Oklahoma line eastward to the Ouachita River, the strike of the basement surface is nearly east-west, and the slope is toward the south at the rate of from 50 to 75 ft. per mile. East of the Ouachita River the strike changes to northeast and continues in that direction to the northeastern corner of the state. The slope over that range is toward the southeast at 50 to 100 ft. per mile almost to the Mississippi River, where the direction changes to southwest. This change in dip coincides with the axis of the upper Mississippi embayment, which is closely paralleled by the course of the Mississippi River.

A local reversal of the southward trend, however, has been deduced by Spooner from wells in the extreme southeast part of Arkansas and adjacent Mississippi and Louisiana. In about the vicinity of the arrows on Figure 1 indicating the synclinal character of the Mississippi embayment, this interpretation places the depth to the basement at from 3,000 to 3,500 ft. below sea level, with a gentle slope northward and a steep slope toward the west.

The basement floor's rate of inclination changes abruptly along a line drawn from southern Little River County (just south of the 2,500-ft. contour on Figure 1) eastward into east-central Ouachita, where a fault is shown, thence northeasterly into Cleveland County. To the north of this line, the basement floor lies at depths ranging from sea level to 2,500 and 3,500 ft. below sea level; to the south, wells hundreds of feet deeper failed to reach the base of the Lower Cretaceous. This abrupt change in the rate of inclination of the basement floor is again noted in Ashley County near the southeastern corner of Arkansas, and is traceable in a general southeasterly direction along the east side of the Monroe and Richland fields in northeastern Louisiana.

Spooner concluded that the abrupt change in the slope of the basement floor may be due in part to sharp flexing and in part to faulting. In a general way the flexure coincides with the approximate outlines of the Lower Cretaceous base of deposition. The best evidence of faulting is found where the basement is penetrated in Sec. 29, T. 12 S., R. 15 W., but in a well four miles south was not encountered at a depth 500 ft. greater. The fault shown on Figure 1 passes between those wells.

BASEMENT FLOOR ON ATLANTIC COAST

Scattered wells along the Atlantic coast have penetrated the basement complex. Contours based on their data have been included in Figure 1 (Johnson, 1931; Sanford, 1913; Stephenson and Johnson, 1912; Mansfield, 1927). They indicate slopes of from 20 to 65 ft. per mile.

GEOPHYSICAL DATA ON BASEMENT CONFIGURATION

GRAVIMETRIC

The use of gravimetric data to supplement other lines of evidence in substantiating the existence of the Gulf Coast geosyncline has been cited. Their bearing on the attitude of the basement complex is largely qualitative. At the same time, the assemblage of regional indications is consistent and significant even though the possibility of several specific interpretations has been cited by Barton, Ritz, and Hickey (1933).

Thom (1937) has published what he describes as a "semi-detailed gravimetric map" showing variations in the vertical component of gravity in

the Middle States Region, as determined by the United States Coast and Geodetic Survey. The anomalies which it defines are for the most part under exposed Piedmont rocks.

SEISMIC

Seismic methods, if applicable at all, should give more quantitative results in this problem. Their contribution to the Gulf Coast geosyncline problem was negative in the sense that direct evidence of the basement rocks was not believed to have been found at depths of the order of 20,000 ft. in the Houston-Jennings area.

The Cretaceous and Cenozoic sediments of the Atlantic and Gulf Coastal Plains are for the most part poorly lithified, except for certain anhydrite and massive limestone horizons in the Lower Cretaceous. The velocity of seismic waves in a column of alternating sands, shales, and chalky members will in general tend to increase continuously rather than discontinuously downward as compaction by pressure affects it. It is well known that the travel-time graph in such a region then becomes a continuous curve. Older rocks, such as Triassic or Paleozoic, tend by advanced lithification to become more sharply differentiated one layer from another, with each layer approaching a condition of nearly constant velocity determined by its composition and history more than by present depth of burial.

SEISMIC VELOCITIES IN THE BASEMENT COMPLEX

Paleozoic rocks of the types presumably present in the basement complex of the Coastal Plain transmit the fastest seismic waves with velocities of from about 15,000 to 20,000 ft./sec. A velocity of 20,000 ft./sec. has been observed in the Pottsville sandstone of Pennsylvanian age in West Virginia (Leet, 1938). Information on Archean rocks is lacking. Triassic lava flows might be expected to fall in about the same range as the Paleozoic rocks. There is every reason to anticipate fluctuations in the velocity from place to place in the basement complex.

SEISMIC VELOCITIES IN THE YOUNGER SEDIMENTS

In the commercial work of the past 15 years, many values have been obtained for velocities in the sedimentary cover of the Gulf Coastal Plain. One method, applied as a basis for reflection work, has been to lower a seismometer to different depths in a well and record directly the travel times from the surface to those depths. This technique, in a column of Cretaceous and Tertiary sediments in southern Arkansas, about 70 miles from the nearest border of the Coastal Plain, gave the results listed in Table 1.

Refraction profiles in Cenozoic sediments of southern Louisiana, showed conditions near the surface to be similar to those represented in Table I. In a specific instance (Rosaire and Lester 1932) velocities under the waters of a bay of the Gulf of Mexico started at 5,000 ft./sec. and increased at the rate of 1.25 ft./sec. per foot of depth down to approximately 2,000 ft., the maximum penetrated by the observations. Profiles

TABLE 1—*Seismic longitudinal velocities in a Cretaceous-Tertiary column 70 miles from the inner margin of the Gulf Coastal Plain in southern Arkansas*

Depth in Ft. below the Surface	Velocity in Ft./Sec.
Surface	5,500
1,000	6,790
2,000	8,080
3,000	9,380
4,000	10,670
5,000	11,950

Average rate of increase: 1.29 ft./sec. per ft. of depth.

of greater length and correspondingly increased penetration in the same region indicated that the rate of increase was slower at greater depths, leading to a value of 11,400 ft./sec. at a depth of 10,500 ft.

Profiles 42,250 ft. in length in southern Texas over the axis of the postulated Gulf Coast geosyncline penetrated 10,700 ft., at which depth the velocity was 12,100 ft./sec.

Gutenberg, Wood, and Buwalda found velocities in Tertiary sediments of the Los Angeles Basin ranging from 6,200 ft./sec. at a depth of 65 ft. to 11,480 ft./sec. at a depth of 5,400 ft.

POSSIBLE DEPTH LIMITATIONS OF SEISMIC METHOD

If an average velocity of 17,500 ft./sec. for the basement rocks is assumed, some idea of possible limitations of the refraction and reflection methods in this problem can be considered. At one extreme, by extrapolating the Arkansas Cretaceous-Tertiary velocities of Table 1, it is found that before a depth of 10,000 ft. is reached velocities in the sediments will equal those in the basement and the discontinuity becomes indeterminate by any method relying upon velocity differences. At the other extreme, this limiting velocity apparently would not be reached in the southern Gulf Coast region at depths less than 20,000 to 25,000 ft. It is possible that in both of these estimates the limits are shallow, because regardless of the specific law of velocity increase in unconsolidated sediments there is a general tendency for the rate to become slower at depth.

TESTS OF SEISMIC METHODS

In 1931 Gutenberg, Wood, and Buwalda, assisted by Henry Salvatori, conducted an extensive series of tests of seismographic methods for deter-

mining crustal structure under a variety of conditions, some of which apply to the Coastal Plain problem. Their conclusions are in part, as follows:

"The velocities of surface waves and of direct compressional waves through the surface parts of the crust can be measured with a high degree of accuracy. . . . By the reflection method, depths to boundary surfaces between rock masses or strata possessing different physical properties can be measured down to 10,000 ft. and, under the most favorable conditions, probably down to 20,000 ft. or more. . . . By the refraction method velocities not merely in the superficial but also in the deeper layers can be accurately determined. . . . Depths to marked boundary surfaces can be computed. . . . The existence of a fault can be proved even if concealed, and the amount of its displacement inferred if conditions are favorable, but it is probably more difficult to determine its dip by the refraction than by the reflection method."

A high-velocity layer (10,170 ft./sec.) under 360 ft. of a lower-velocity layer (5,400 ft./sec.) gave rise to a very sharp later phase which sufficient observations demonstrated was a wave reflected from the discontinuity.

SEISMIC INVESTIGATIONS ON THE ATLANTIC COASTAL PLAIN

A project was organized in 1935 to develop and apply a seismic method for securing

"exact information in regard to the basement of the Coastal Plain sediments, both in the emerged and in the submerged portions, from the inner margin at the border of the Piedmont Plateau to the edge of the Continental Slope, to determine the thickness of the Coastal Plain sediments and, if possible, some facts concerning their character."

Grants by the Geological Society of America from the Penrose Bequest made it possible to undertake the projects, with cooperation from a number of other sources, including fellowships from the Guggenheim Foundation and the American Geophysical Union. Three reports have been published to date, Part I by Ewing, Crary, and Rutherford (1937), Part II by Miller (1937), and Part III by Ewing, Woollard, and Vine (1939).

CAPE HENRY, VIRGINIA, SECTION

Part I (1937) describes refraction measurements at 15 stations on a line approximately 150 miles long, crossing the entire Coastal Plain at Cape Henry, Virginia, (about 76° West Longitude, 37° North Latitude on Fig. 1). According to Part II (p. 806)

"The line chosen was particularly favorable, because reliable records of two deep wells, drilled to the underlying crystalline rocks near the mouth of Chesapeake Bay, were available. This permitted Ewing and his associates to determine rates of vibration propagation in known rocks before entering the entirely unknown region. In this way, a valuable check was possible."

The land stations constituted an incomplete but relatively satisfactory test of adaptability of the method for determining depths to buried crystalline rocks of the basement complex under shallow cover. The shooting designed to obtain velocity calibrations on shore and then carry the work into unknown territory at sea, however, met with serious difficulties which

do not seem to have been appreciated where the results have been quoted in Part II and elsewhere (Heck, 1937 and 1938; Thom, 1937 p. 317).

To determine the true velocity of seismic waves in a buried formation, it is necessary to make two sets of observations at a given station, arranged in what are called "reversed profiles." (Leet, 1938a, Ewing, 1937, p. 764, or any discussion of refraction seismic profiles.) On the Cape Henry section, this was done at only one of the 15 locations. At certain other places, the instruments were left at their stations while shots were fired at successively greater distances in a manner used and discussed by Gutenberg (1931). This gives an offset in the high-velocity line in the presence of considerable slope of the discontinuity, or a fault. At two of the locations, this technique was so applied as to leave reasonable assurance that no large slope was present in the basement rocks. Elsewhere on this section, true velocities were not observed and quantitative information on lithologic changes inferred from velocity changes was therefore lacking.

Neither true velocity determinations nor tests for slope were made at the sea stations, the outermost of which was approximately 90 miles from the control well at which calibrations were to have been made.

CALIBRATION

Part I (Ewing *et al.* 1937) described the calibration profile at Ft. Monroe as "a poor measurement." Various conditions

> "made it impossible to get seismograms which could be read with high accuracy. Further, only one shot was made, and the data are, therefore, meager and do not suffice to give velocities in the semi-consolidated and crystalline zones at this point."

SEA STATIONS

The authors of Part I pointed out that they

> "consider the present work the initial application of seismic measurements to submarine geological problems."

Further, the sea stations represent an all-important step into unknown territory, since the work on land was essentially a check controlled by well data and did not change previous pictures of the basement configuration. The chief interest in the investigation from the standpoint of new geological information rests on the sea stations.

Several reflection shots were made at sea, but the authors of Part I stated that,

> "the stations are so far apart that correlation of the reflecting horizons is not possible,"

and concluded that,

> "As no reflections were found that could be identified with the surface of the crystalline, it will probably not be possible to map the crystalline in this region by reflections."

If the basement rock velocities are not sufficiently differentiated from those of the cover to give recognizable reflections, there is cause for serious doubts as to the ability of the refraction method to map the discontinuity.

There were 4 sea stations, 8, 12, 16, and 20. Depths at 8 and 12 indicated a tendency for the basement slope to steepen, so the crucial points were at stations 16 and 20, nearest the 100-fathom break in seabottom slope. At Station 16, a record was obtained at each of two points on a line 39,500 ft. long. The authors of Part I stated,

"This determination is incomplete, because both the number of shots and the amount of explosive used on each shot had to be curtailed because of shortage of the dynamite supply."

A minimum depth for the crystallines was estimated, however, and the computed profile of the basement broken at that point, to indicate that basement rocks were not picked up, according to the authors' interpretation.

The profile at Station 20, where the average depth of water was 35 fathoms, thus stands alone in support of the concluding statement of Part I (p. 801),

"The thickness of unconsolidated and semi-consolidated material (velocity of longitudinal waves about 8,000 ft./sec.) near the edge of the Continental Shelf was found to be about 12,000 ft."

For this profile, 44,000 ft. in length, there were also only two points as at Station 16. Here it was stated, however, that

"the data are more complete than at Station 16 but are still decidedly inadequate, because of the shortage of the explosive supply."

The second of the two points was assigned to the crystallines and the first to the "semi-consolidated". Lines through these were then drawn with slopes corresponding to velocities obtained by extrapolation from other stations, though the nearest at which a true crystalline velocity had been actually observed was nearly 100 miles westward. Depths were computed with the conclusion that

"the depth to the crystalline zone obtained here is, therefore, reliable insofar as the basic assumption—that the three zones revealed by the stations to the west persist here—is justified, for which there is considerable evidence from the seismograms."

This, of course, meant assuming a fundamental part of the answer which was being sought.

Since the data were admittedly inadequate and required a choice of assumptions for their interpretation, consideration of alternative assumptions might not be out of order. The points of Station 20 have been plotted in Figure 2 on a graph of a normal profile in southern Gulf Coast sediments. From the Gulf Coast profile, it has been computed that the ray reaching 42,000 ft. penetrated about 10,700 ft. into sediments, at

which depth the velocity was 12,100 ft./sec. The comparison emphasizes primarily the importance of control between the origin and such distant points, and to some extent the range of uncertainty in interpreting the two isolated points of Station 20. The possibility that Station 20 did

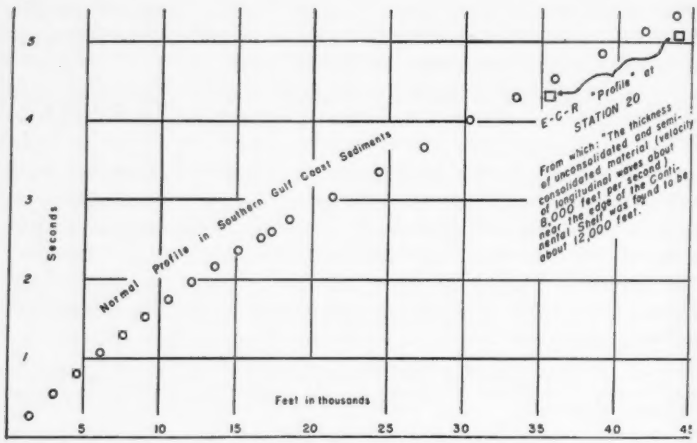
Seconds
Normal Profile in Southern Gulf Coast Sediments
E-C-R "Profile" at STATION 20
From which: "The thickness of unconsolidated and semi-consolidated material (velocity of longitudinal waves about 8,000 feet per second) near the edge of the Continental Shelf was found to be about 12,000 feet."
Feet in thousands
5
10
15
20
25
30
35
40

FIGURE 2.—*Comparison of Sea Station 20 with normal profile in Gulf Coast sediments*

Circles give time-travel data for normal profile in Gulf Coast sediments. Squares give time-travel data obtained by Ewing, Crary, and Rutherford for their station 20.

not pick up the crystallines at all is fundamental from the standpoint of the geological significance of the results.

BARNEGAT BAY, NEW JERSEY, SECTION

Report III describes investigations between Princeton, N. J., and Barnegat Bay (Ewing, Woollard, and Vine, 1939). In this, Woollard presents a detailed analysis of well-log and other geologic data pertaining to the general problem. Eight refraction profile stations were established in the seismic work. Five of these were essentially tests in regions controlled by well data, though the authors

"wish to stress one point in particular: that the seismic results were calculated without the benefit of any of the detailed geologic data later used for correlation. These data, all from unpublished well records, were not consulted until after the seismic results had been submitted to the New Jersey State Department of Geology to see if any correlation existed between the seismic and geologic horizons."

The other 3 profiles extended depth information on basement rocks to the coast line, with no marked change in rate of seaward dip observed.

A buried fault was indicated by one shot on the profile at Plainsboro but no further shooting was done to detail it or eliminate other possible interpretations of the seismic data, which the authors suggested. There appears to be a difference in the direction of throw indicated by the seismic data and that inferred from geologic evidence in the vicinity (Part III, p. 275 and Fig. 16, p. 290).

The region covered by the investigations of Part III is of special interest at the present time owing to the fact that it includes the epicentral zone of a series of earthquakes on August 22, 23, and 27, 1938 (Anonymous, 1938).

CONCLUSIONS

The present position of the basement rocks underlying the Atlantic and Gulf Coastal Plain is known in many places from the logs of wells which penetrated them. Deep wells and geophysical data indicate the presence of a geosyncline which carries basement rocks to depths greater than 35,000 ft. near the margin of the Gulf of Mexico. Corresponding information about the configuration of the basement surface along the Atlantic coast has been sought by geophysical methods, notably the seismic.

Data from commercial seismic prospecting point to the possibility that sufficient velocity differences for successful application of the seismic method to mapping basement rocks may not exist if the cover of sediments has a thickness of between 10,000 and 20,000 ft., or if the basement is overlain by such high-velocity formations as massive Cretaceous limestone or anhydrite.

Seismic refraction observations at sea off Cape Henry, Virginia, were made by Ewing, Crary, and Rutherford (1937), who reported sediments 12,000 ft. thick covering basement rocks about 150 miles downdip from where basement rocks disappear under the sedimentary cover. It has been shown that that computation, based upon records from one station 35,500 ft. and one station 44,030 ft. from a single shot, relies upon assuming structural conditions which were being investigated and upon seismic travel-time measurements so ambiguous as to give no assurance that basement rocks were penetrated at all.

WORKS TO WHICH REFERENCE IS MADE

Barton, D. C., Ritz, C. H., and Hickey, Maude (1933) *The Gulf Coast geosyncline,* Am. Assoc. Petrol. Geol., Bull., vol. 17, p. 1446-1458.

Billings, M. P., and Williams, Charles (1932) *Origin of the Appalachian Highlands,* Appalachia, vol. 19, p. 1-33.

Darton, N. H. (1938) *Structural relations and history of Cretaceous and Tertiary formations of the central Atlantic Coastal Plain,* abstract, Geol. Soc. Am., Pr. 1937, p. 76.

Ewing, Maurice, Crary, A. P., and Rutherford, H. M. (1937) *Geophysical investigations in the emerged and submerged Atlantic Coastal Plain, Part I, Methods and results,* Geol. Soc. Am., Bull., vol. 48, p. 753-802.

———, Woollard, G. P., and Vine, A. C. (1939) *Geophysical investigations in the emerged and submerged Atlantic Coastal Plain, Part III, Barnegat Bay, New Jersey, Section,* Geol. Soc. Am., Bull., vol. 50, p. 257-296.

Gutenberg, B., Wood, H. O., and Buwalda, J. P. (1932) *Experiments testing seismographic methods for determining crustal structure,* Seis. Soc. Am., Bull., vol. 22, p. 185-246.

Heck, N. H. (1937) *Comments before International Committee on Ocean Basins,* Edinburgh. Mimeographed for private distribution.

——— (1938) *The role of earthquakes and the seismic method in submarine geology,* Am. Phil. Soc., Pr., vol. 79, p. 101-102.

Johnson, M. E. (1931) *The mineral industry of New Jersey for 1929,* N. J. Dept. Conserv. Devel., Geol. Ser., Bull. 36, p. 28.

Leet, L. D. (1938a) *Practical seismology and seismic prospecting,* D. Appleton-Century Co., N. Y.

——— (1938b) *Longitudinal velocities in some weathered and unweathered Carboniferous rocks,* Seis. Soc. Am., Bull., vol. 28, p. 163-168.

Mansfield, W. C. (1927) *An oil-prospecting well near Havelock, North Carolina,* N. C. Dept. Conserv. Devel., Econ., Paper 58.

Miller, B. L. (1937) *Geophysical investigations in the emerged and submerged Atlantic Coastal Plain, Part II: Geological significance of the geophysical data,* Geol. Soc. Am., Bull., vol. 48, p. 803-812.

Rosaire, E. E., and Lester, O. C., Jr. (1932) *Seismological discovery and partial detail of Vermilion Bay salt dome, Louisiana,* Am. Assoc. Petrol. Geol., Bull., vol. 16, p. 1226.

Sanford, Samuel (1913) *The underground water resources of the Coastal Plain province of Virginia,* Va. Geol. Surv., Bull. 5, p. 100-101.

Spooner, W. C. (1935) *Oil and gas geology of the Gulf Coastal Plain in Arkansas,* Ark. Geol. Surv., Bull. 2.

Stephenson, L. W. (1928) *Marine transgressions and regressions of the Atlantic Coastal Plain,* Amer. Jour. Sci., 5th ser., No. 94, p. 281-298.

———, and Johnson, B. L. (1912) *The Coastal Plain of North Carolina,* N. C. Geol. Econ. Surv., vol. 3, p. 342.

Stetson, Henry C. (1936) *Geology and paleontology of the Georges Bank canyons,* Geol. Soc. Am., Bull., vol. 47, p. 339-366.

Thom, W. T., Jr. (1937) *Position, extent, and structural makeup of Appalachia,* Geol. Soc. Am., Bull., vol. 48, p. 315-322.

(1938) *New Jersey shocks of August 22, 23, and 27, 1938.* Seis. Soc. Am. East. Sec., Earthquake Notes, vol. 10, p. 2. (Anonymous.)

Harvard University, Cambridge, Mass.

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PETROGRAPHIC PROVINCE OF CENTRAL MONTANA

BY ESPER S. LARSEN

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ABSTRACT

The petrographic province of Central Montana is arbitrarily defined to include the Tertiary igneous rocks near the eastern flanks of the Rocky Mountains from the Yellowstone National Park on the south to the Canadian boundary on the north (Fig. 1).

Each mountain range is made up of one or more groups or subprovinces of igneous rock. The Yellowstone Park area contains at least three subprovinces, the Highwood Mountains four, and the Bearpaw Mountains about seven. The rocks of a subprovince may vary from mafic to felsic, but they are all related in mineral composition, and their chemical analyses fall near smooth variation curves. The mafic rocks of any subprovince differ from those of other subprovinces more than do the felsic rocks. The latter tend toward a granite of about the same composition for all the subprovinces. The mafic rocks of the various subprovinces range from ordinary basalts or gabbros to orthoclase gabbro, to plagioclase shonkinite, to shonkinite without plagioclase. In general, if the mafic rocks of a subprovince are rich in plagioclase all the rocks of that subprovince are rich in plagioclase whereas if the mafic rocks are poor or lacking in plagioclase all the rocks of that subprovince are poor or lacking in plagioclase.

The parent magmas of the various subprovinces are believed to be represented by the mafic rocks. The gradational character of these from gabbro to shonkinite leads to the conclusion that they have a common origin. They are believed to have been formed by the slow differentiation at great depth of a basaltic magma by the separation of plagioclase and hypersthene with some olivine.

From time to time parts of this changing magma were erupted toward the surface where they further differentiated to form the rocks we see. This latter differentiation was relatively rapid and was by crystal fractionation, modified in all the magmas except the shonkinitic magma by assimilation of granitic material.

INTRODUCTION

The petrographic province of Central Montana has been described by Pirsson (1905b), but since then much new information about this province has become available, and the science of petrology has advanced. The present paper is an outgrowth of a field and laboratory study of the igneous rocks of the Highwood Mountains, reconnaissance trips into some of the neighboring mountains, and a study of the literature.

The Tertiary igneous rocks of Central Montana well illustrate the facts that a petrographic province must be defined in respect to both time and space and that it may be made up of a group of subprovinces. The vari-

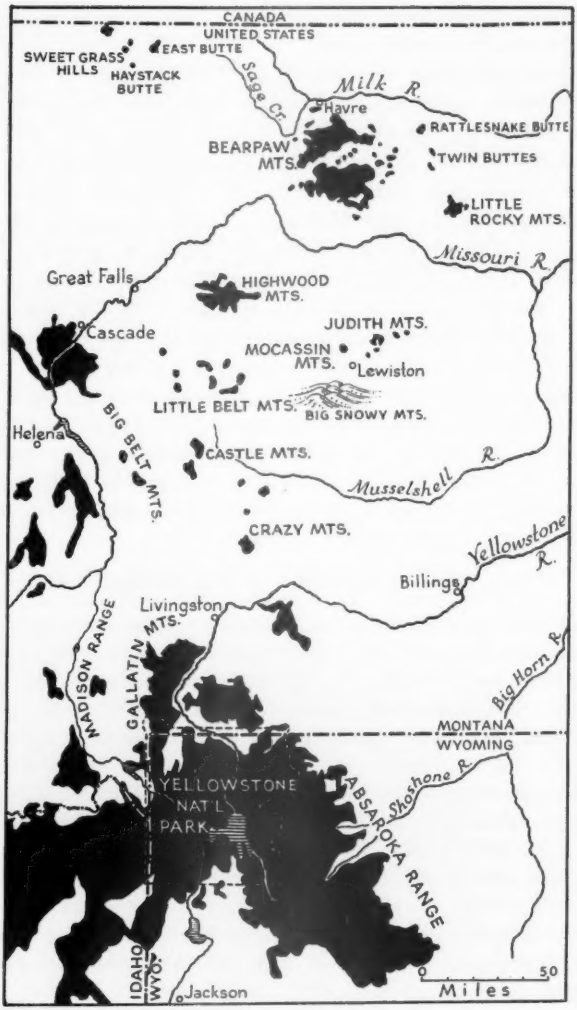

FIGURE 1.—*Sketch map of parts of Montana and Wyoming*
Showing the mountain ranges of Tertiary igneous rocks.

ous rocks of each subprovince are closely related but are less closely related to the rocks of the other subprovinces. The relations of the rocks to structural features and the method of extrusion and intrusion are as

characteristic of the subprovince as are the chemical, mineralogical, and textural peculiarities of the rocks.

The Central Montana petrographic province is a part of the group of Tertiary igneous rocks that are widely distributed along the eastern flanks of the Rocky Mountains (Fig. 1). The chief areas that contain igneous rocks that are assigned to the petrographic province of Central Montana, named from south to north, are: Yellowstone National Park and Absaroka Range, Crazy Mountains, Castle Mountains, Little Belt Mountains, Judith Mountains, Moccasin Mountains, Highwood Mountains, Little Rocky Mountains, Bearpaw Mountains, Northern end of Big Belt Mountains, and the Sweetgrass Hills. The Big Snowy Mountains are probably underlain by a laccolith which is not exposed.

Many of the mountains contain a variety of igneous rocks, and some of the larger areas of igneous rocks are made up of two or more distinct subprovinces or successions of igneous rocks. The extremes of these subprovinces are the lime-alkalic rocks and the alkalic group made up of shonkinite, nepheline syenite, phonolites, and related rocks. There is more than one group of alkalic rocks, and there are groups of rocks intermediate between the two extreme groups.

GENERAL CHARACTER OF THE SUBPROVINCES

The various subprovinces of Montana range from about average lime-alkalic rocks to the potash-rich plagioclase-free leucite rocks of the Highwood and Bearpaw Mountains. The mafic rocks characterize a province better than the felsic rocks, since the extreme felsic rocks of all the subprovinces that have quartz-bearing rocks tend to be granites and rhyolites.

At one extreme are the lime-alkalic rocks of the Crazy Mountains stocks and the lavas, ranging from rhyolite to basalt, in and near Yellowstone National Park. The mafic rocks of these areas are basalts and gabbros and are made up chiefly of calcic plagioclase, augite, hypersthene, and olivine.

In the Absaroka range and elsewhere in and near the Yellowstone National Park, another group of rocks has trachydolerites and orthoclase gabbros as its mafic rocks. Trachydolerites also form a large mountain mass west of Cascade, at the northern end of the Big Belt Mountains. Mineralogically the trachydolerites carry considerable orthoclase and little or no hypersthene but are otherwise much like the basalts. They carry occasional leucite or other feldspathoids. The quartz latites which make up the older volcanic rocks of the Highwood Mountains probably belong to a trachydolerite subprovince.

The subprovince of the Castle and Little Belt Mountains is not far from the trachydolerite subprovince, but the mafic rocks have somewhat more orthoclase than plagioclase and are plagioclase shonkinites. The stock of Highwood Peak in the Highwood Mountains, the stocks of the Sweetgrass Hills, some rocks of the Bearpaw Mountains, and many of the stocks and laccoliths of the mountains in the eastern part of the province are of this intermediate type with both plagioclase and orthoclase.

The alkalic rocks of the northern area of the Crazy Mountains are rich in soda as well as in potash. They contain little plagioclase and have abundant nepheline in some of the felsic syenites.

The younger volcanic rocks and nearly all the intrusive rocks of the Highwood Mountains are rich in potash. The mafic types of this area are mafic phonolites and shonkinites. Plagioclase and hornblende are rare in them, and leucite, analcime, nepheline, or other feldspathoids are abundant. Quartz is absent, and the syenite appears to be the final differentiate. The greater part of the volcanic rocks and some of the intrusive rocks of the Bearpaw Mountains are similar to the Highwood rocks but appear to be even richer in potash.

The small bodies of alnoite and monticellite peridotite in the Highwood Mountains and east of Winnett probably belong to a lime-rich subprovince that may have originated from some special process.

Where the age relations of the different subprovinces have been determined the older rocks are commonly nearer the lime-alkalic types. In the Absaroka range the trachydolerites are both underlain and overlain by lime-alkalic rocks. In the Highwood Mountains the quartz latites are older than the potash-rich rocks and are separated from them by a long interval of erosion. In the Crazy Mountains the lime-alkalic rocks are older than the alkalic rocks. In the Bearpaw Mountains the succession of subprovinces from oldest to youngest is quartz monzonite-syenite subprovince, lime-alkalic subprovince, subprovince of shonkinite with a little plagioclase, monzonite, shonkinite subprovince, and finally leucite-rich subprovince.

Some minerals are characteristic of the subprovinces. Hypersthene is confined to the lime-alkalic rocks. Amphiboles are abundant in the andesitic rocks of the subprovinces with moderate potash content and of the Little Belt types but are subordinate in the alkalic rocks of the Crazy and Highwood types. Plagioclase decreases in amount from the lime-alkalic subprovinces to the alkalic subprovinces and is rare in the Highwood subprovince. In general if plagioclase is abundant in the mafic rocks of a subprovince it continues to be correspondingly abundant in the felsic rock, except in the extreme rhyolites and granites. If it is

in small amount or absent from the mafic rocks, it is in small amount or absent from all the rocks of the subprovince. This is well shown in the differentiated laccoliths and stocks. In the differentiated laccoliths of the Highwood Mountains plagioclase, with the exception of albite, is absent from all the rocks, ranging from shonkinite to nepheline syenite (Hurlbut and Griggs, 1939). In a similar differentiated laccolith near Box Elder in the Bearpaw Mountains (Pecora, personal communication) and in the differentiated stock of Yogo Peak in the Little Belt Mountains (Weed and Pirsson, 1895b), plagioclase is present but subordinate to orthoclase in all the rocks—plagioclase shonkinite to syenite or nepheline syenite. In the differentiated sheet southeast of Electric Peak in the Yellowstone National Park both the mafic and felsic rocks contain plagioclase as the chief felsic mineral (Hague, Iddings, and Weed, 1899, p. 82-83).

In general, the lime-alkalic rocks form stocks with radiating dikes but with few sills or laccoliths, whereas the alkalic rocks, including the trachydolerites, form stocks with radiating dikes and numerous sills and laccoliths. The alkalic types are rich in barium and strontium, and these oxides increase in amount as the alkalies increase. The alkalic rocks are relatively poor in titanium, phosphorus, and zirconium.

The larger mountain masses contain a variety of rocks ranging from highly mafic to highly felsic types. Some of the larger laccoliths contain rocks varying from shonkinite to nepheline syenite and show the course of differentiation of the magma by crystal sorting.

Nearly all the rocks of each of the smaller mountain masses have a close qualitative mineralogical similarity. They may vary from mafic varieties such as shonkinite to felsic syenites or even granites. They form a subprovince which is different from the subprovinces of the other mountains. The larger mountains commonly contain more than one subprovince.

For each subprovince the chief rocks of the lavas, stocks, and dikes are much like those of the differentiated laccoliths. Some of the felsic rocks have minerals, such as amphibole and quartz, that are absent from the mafic rocks. Chemically the rocks of each subprovince fall near a regular variation diagram. They show a clear relationship, and the various rocks must have been derived from a common parent magma by a rather simple process of differentiation.

YELLOWSTONE NATIONAL PARK AREA

The Tertiary volcanic rocks of the Yellowstone National Park and of the Absaroka range to the north and east of the Park have been mapped and described by Hague, Iddings, and Weed (1899), Hague,

and Iddings (1896), Iddings and Weed (1894), and Hague (1899). Later detailed descriptions of smaller areas have appeared.

A generalized section from youngest to oldest of the chief Tertiary igneous rocks as modified from Rouse (1937, p. 1272) is given in Table 1.

TABLE 1.—*Sequence of the chief eruptions in the Yellowstone National Park Area*

Rhyolite	
Widespread erosion	
Warping	Ishawooa intrusives
Late basalt sheets	
Late basic breccia	Sylvan dikes
Erosion	
Late acid breccia	Electric Peak intrusives
Early basic flows	Diorite intrusives
Erosion	Sunlight intrusive
Tilting and folding	
Early basic breccia	Crandall volcano
Local folding and erosion	
Early basic tuff	
Extensive erosion	
Early acid breccia, trachytic rhyolite	

Taking into account the field relations, the petrographic character, and the chemical analyses, most of the rocks of this large area fall into three main groups or petrographic subprovinces, which are characterized as follows:

(1) Yellowstone subprovince. To this subprovince belong the early acid breccia(?), the late acid breccia and all the later extrusive rocks, the Ishawooa intrusive rocks, the Haystack stock (Emmons, 1908a), and most of the other large intrusive bodies except the Sunlight intrusive rocks and the Crandall volcano. The Yellowstone is a lime-alkalic subprovince made up of rocks ranging from basalt to rhyolite and their intrusive equivalents. Sixty-three analyses of rocks from this subprovince are available. They are plotted on a variation diagram (Fig. 2) of the type proposed by Larsen (1938). Most of the analyses fall near regular variation curves. These analyses are plotted on a norm diagram in Figure 12. The Yellowstone rocks as compared with the lavas of the San Juan Mountains which are selected as a typical, well-studied group, as shown in Figure 5, are low in K_2O, especially at the felsic end, high in MgO and low in FeO at the mafic end, and high in SiO_2 at the felsic end.

(2) Crandall subprovince. The 16 analyzed rocks of the Crandall volcano (Hague, Iddings, and Weed, 1899, p. 260-261) belong to this subprovince. All the rocks of the early basic breccia and the trachytic rhyolite may belong here. One of the analyzed specimens from the

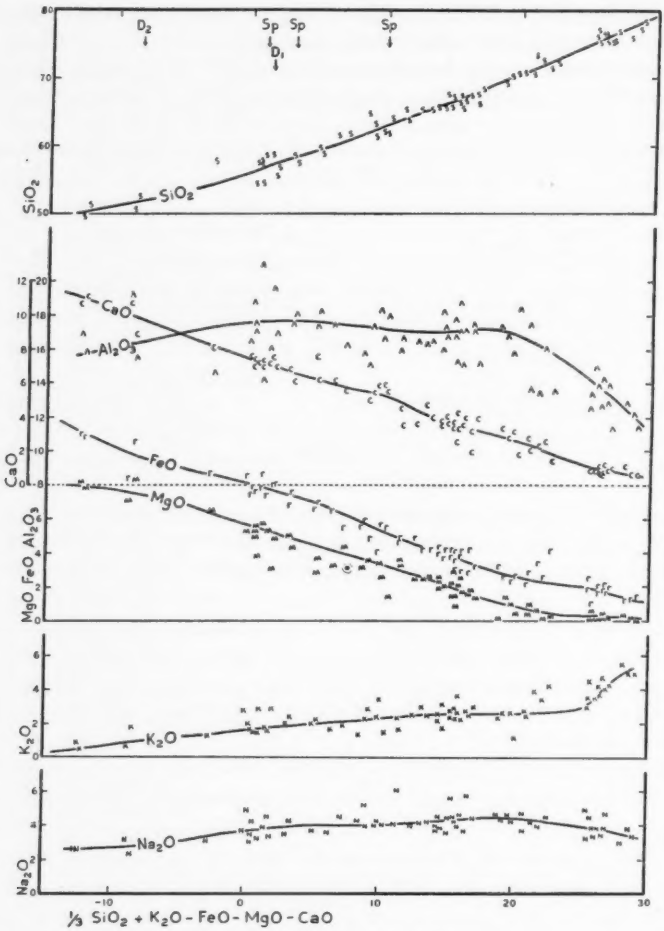

FIGURE 2.—*Variation diagram of rocks of the Yellowstone subprovince*
D = Lower acid breccia. S_p = Early basic breccia of Sepulchre Mountain.

Crandall volcano is from a basalt flow in the early basic breccia, one from a fragment of the breccia, and the other 14 from intrusive bodies which include gabbro, orthoclase gabbro, diorite, quartz diorite, monzonite, and aplite. A variation diagram of these rocks is shown in Figure 3. The analyses fall near regular curves.

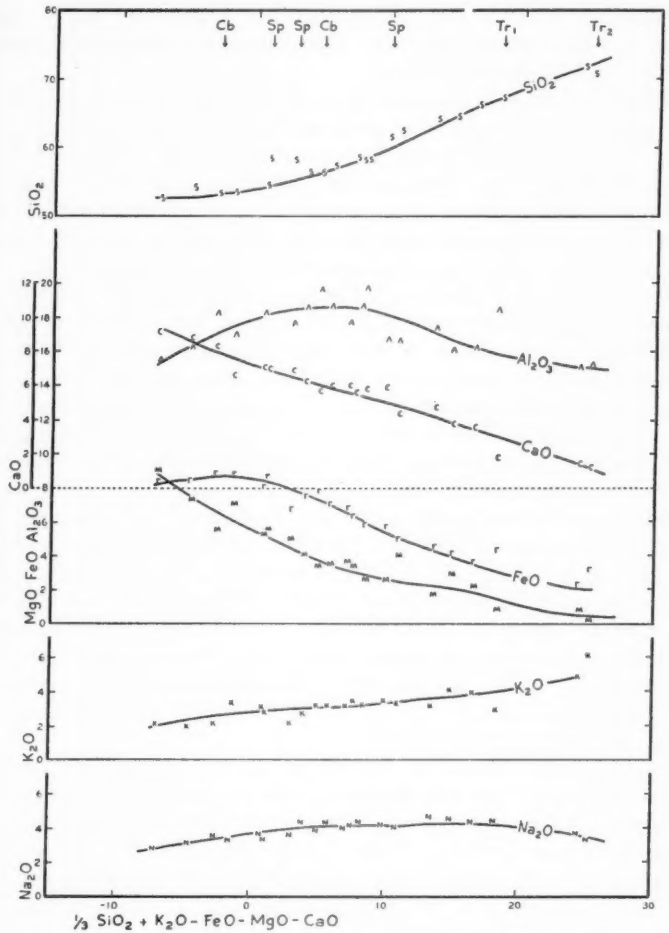

FIGURE 3.—*Variation diagram of rocks of the Crandall subprovince*

Cb = Early basic breccia of Crandall area.
Sb = Early basic breccia of Sepulchre Mountain area.
Tr = Trachytic rhyolite.

Two specimens from the lower basic breccia of the Crandall area and three from Sepulchre Mountain have been analyzed. In Figures 2 and 3 these analyses are marked *Cb* for the Crandall area and *Sp* for the Sepulchre Mountain area. The rocks from the latter area are plotted on

variation diagrams for both the Crandall and Yellowstone areas. They do not fit either diagram very well. The dikes of the Sepulchre Mountain area fit the Yellowstone diagram.

Iddings and Weed (1894, p. 321) state that the trachytic rhyolite is intimately associated with the early acid breccia. The two analyses of rocks from the trachytic rhyolite are plotted on the diagram for the Crandall subprovince. They are different from those of the Crandall volcano though nearer to them than to those of the Yellowstone subprovince. One analysis is somewhat like the rocks of the Yellowstone subprovince; the other is nearer those of the Crandall subprovince.

The rocks of the Crandall subprovince as compared with those of the Yellowstone subprovince are from 1 to nearly 2 per cent higher in K_2O and over 2 per cent lower in SiO_2 over most of the range. They are somewhat like the lavas of the San Juan Mountains but are a little higher in K_2O and MgO and lower in FeO and SiO_2 for the mafic rocks. A triangular diagram of the norms of these rocks is shown in Figure 14.

A group of intrusive bodies from the northern slopes of the Beartooth Mountains have recently been described by Rouse, Hess, Foote, Vhay, and Wilson (1937) as quartz monzonite, keratophyre, and dacite porphyries. Five analyses are given. The rocks must be considerably altered as they contain from 0.55 to 2.64 per cent CO_2. They are all siliceous and vary greatly in their content of soda. When plotted on a variation diagram, the analyses do not fall near a regular curve. As compared with the rocks of the Yellowstone subprovince, they are low in SiO_2 and FeO and high in alkalies and Al_2O_3. In silica content they are nearer the rocks of the Crandall subprovince, but they are higher in Al_2O_3 and Na_2O and lower in FeO and K_2O.

(3) Absaroka subprovince. To this subprovince belong the early basic flows (or sheets) and the Sunlight group of intrusives. Near Valley, Wyoming, in the southeastern part of the Ishawooa quadrangle, the early basic flows were divided by Rouse (1933; 1935; 1937) into six divisions. Rouse gives analyses of six rocks from three of these divisions, and they probably all belong to the Absaroka subprovince, although the rocks from zones 3 and 5 are low in K_2O and may belong to the Crandall subprovince (marked V_3, V_5 in Fig. 4). The andesite dike and the quartz diorite porphyry of the Valley area belong to the Ishawooa intrusive rocks of the Yellowstone subprovince. Parsons (1939) has described the volcanic centers of the Sunlight area, in the Crandall quadrangle, and 13 analyses of intrusive rocks from this area are available. The Sunlight stock belongs to the Absaroka subprovince as do all the dikes. One of the gabbros from the White Mountain stock is rather different.

The rocks of the Absaroka subprovince are chiefly trachydolerite, orthoclase gabbro, diorite, monzonite, syenite, and trachyte. A variation dia-

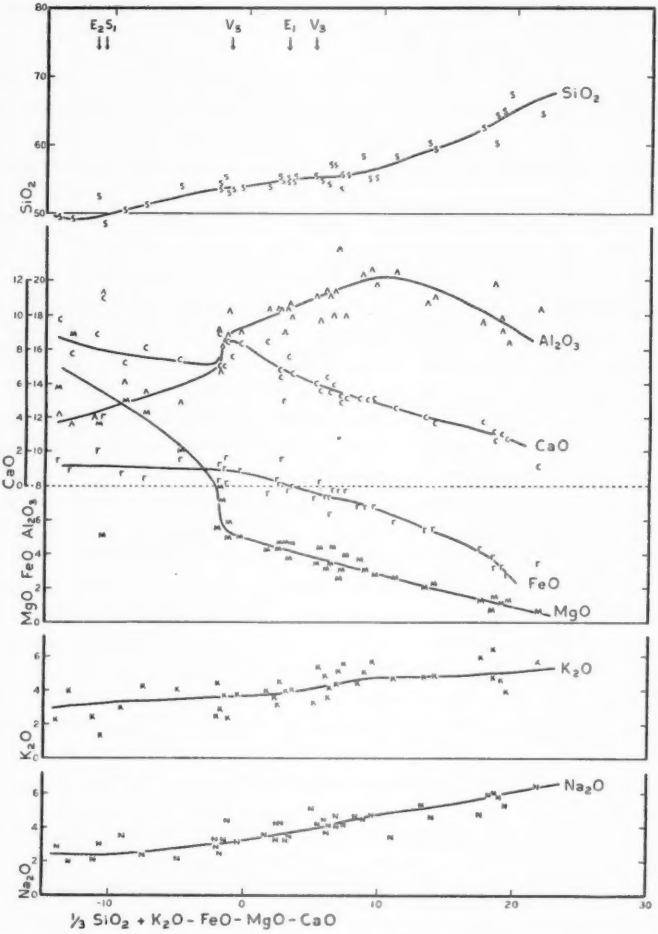

FIGURE 4.—*Variation diagram of rocks of the Absaroka subprovince*
E = Differentiated sill on the northeast shoulder of Electric Peak.
S_2 = White Mountain stock.
V_3 = Zone 3 of Valley area. V_5 = Zone 5 of Valley area.

gram of 39 analyses representing this subprovince is shown in Figure 4, and a triangular diagram of the norms in Figure 15. The analyses show high K_2O, low CaO and SiO_2, and, for the felsic rocks, high Na_2O.

The differentiated sheet on the southeast shoulder of Electric Peak (Hague, Iddings, and Weed, 1899, p. 83) is rich in K_2O. Analyses of the

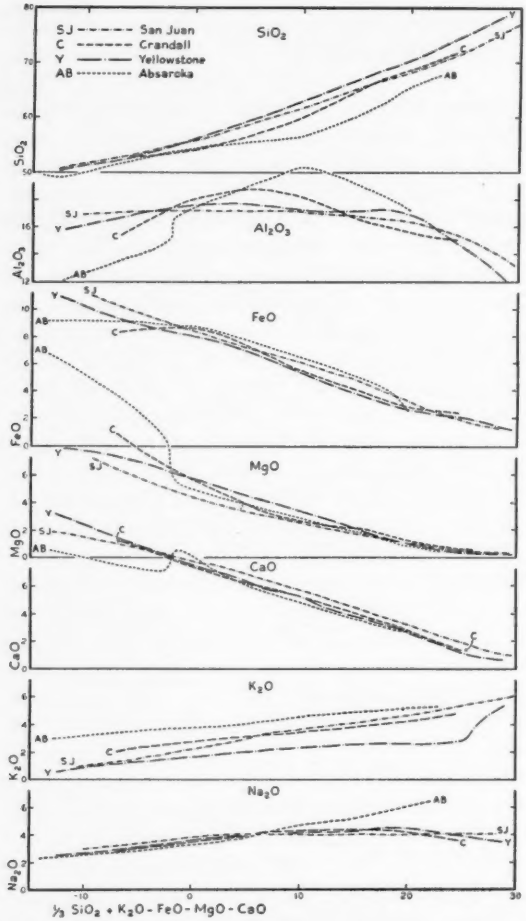

FIGURE 5.—*Variation curves for the three petrographic subprovinces* Yellowstone area and the province of the San Juan Area, Colorado.

augite andesite porphyry of the upper part (E_1) and of the pyroxene-enriched lower part (E_2) are plotted on the diagram for the Absaroka subprovince. They probably belong to a related province.

Figure 5 shows variation curves for the three subprovinces of the Yellowstone area and, for comparison, those for the lime-alkalic subprovince of the San Juan lavas of Colorado. In the Yellowstone Park area

the succession of eruptions by petrographic provinces appears to have been:

First, rocks of the Yellowstone subprovince (?) (early acid breccia)
After an interval of erosion the rocks of the Crandall subprovince were erupted (early basic breccia)
After another interval of erosion the rocks of the Absaroka subprovince were erupted (early basic flows)
The main rocks of the Yellowstone subprovince were then erupted.

CRAZY MOUNTAINS

Wolff (1938) has given an excellent description of the igneous rocks of the Crazy Mountains. He describes an older lime-alkalic series, chiefly in the southern part of the mountains and making two stocks and associated dikes, with few sills and laccoliths, and a younger alkalic series, found chiefly in the northern part of the mountains and occurring chiefly as sills, laccoliths, and dikes.

The stocks are chiefly diorite and contain a small body of granodiorite and minor bodies of gabbro and picrite. A small amount of quartz and orthoclase is present in the diorite. Hornblende, biotite, and augite are the chief dark minerals. Hypersthene is present in some of the rocks. The smaller intrusives range from aplite and granite porphyry to porphyritic diorite and lamprophyres. Six analyses of rocks from the stocks and seven of those from the dikes are given by Wolff. They are plotted on a variation diagram in Figure 6 and on a triangular diagram of the norms in Figure 16. The analysis at position +3.0 on the variation diagram does not fit the curve. It is Wolff's specimen 343a'89 and is described as coming from near the contact.

The analyzed gabbro and picrite are unusually rich in mafic minerals and were probably formed by accumulation of augite and olivine crystals. These analyses fall to the left of the diagram and are not shown. The rocks of this group in the Crazy Mountains fall fairly near the variation curves for the San Juan province, except that they are a little low in FeO.

The bodies of the northern part of the Crazy Mountains are chiefly alkalic and are richer in soda than any of the other groups of rocks of the petrographic province of Central Montana. They form a distinct subprovince which will be called the Crazy Mountains subprovince. These rocks include granite porphyry, syenite, nepheline syenite, shonkinite (called theralite by Wolff), and lamprophyres. Plagioclase and hornblende are present in a few of the types but are much less common than in those of the Little Belt area. The shonkinites have nepheline

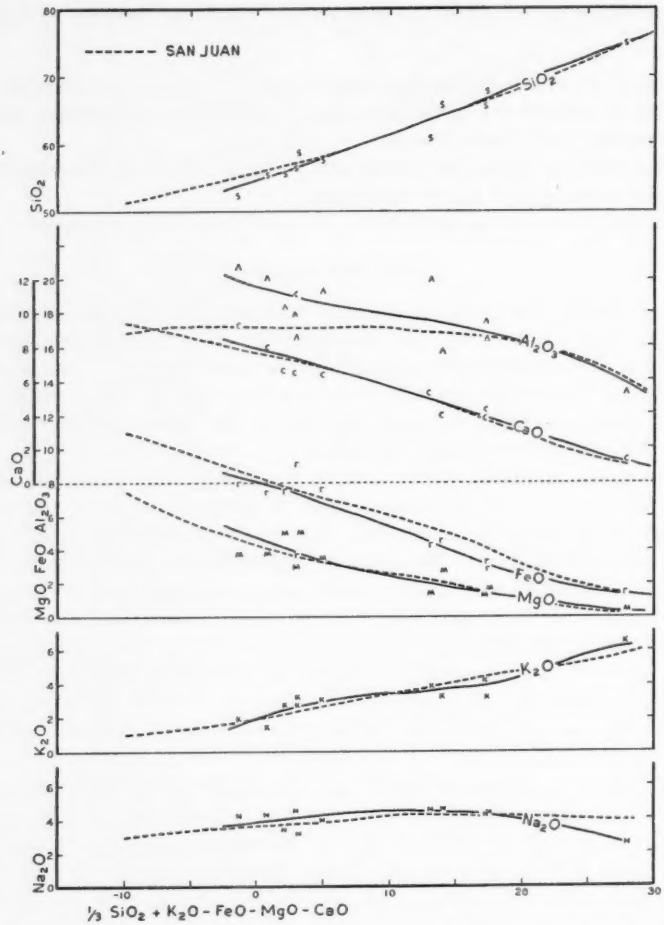

FIGURE 6.—*Variation diagram of rocks of lime-alkalic subprovince of the Crazy Mountains*

With curves for the San Juan, Colorado, province for comparison.

or sodalite and rare hornblende in contrast with the somewhat similar rock of Yogo Peak in the Little Belt Mountains, which has much plagioclase and no nepheline.

A variation diagram for this subprovince is shown in Figure 7 and a triangular diagram of the norm in Figure 17. The soda content is vari-

able but is higher than in the rocks of the other subprovinces of Central Montana. The analyses are fairly high in K_2O and low in FeO and SiO_2.

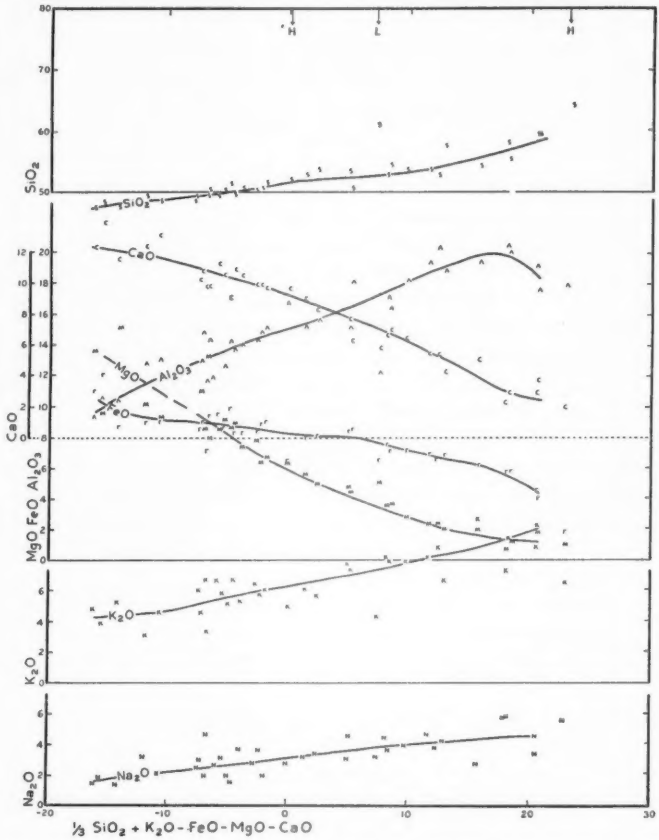

FIGURE 7.—*Variation diagram of rocks of the Crazy Mountains subprovince*

CASTLE AND LITTLE BELT MOUNTAINS

The Castle and Little Belt Mountains are adjoining ranges, and the rocks of the two fall on a single variation diagram. In both mountains there are many small igneous bodies. In the Castle Mountains (Weed and Pirsson, 1896c) granite and granite porphyry predominate; a small

stock of diorite and small bodies of syenite and lamprophyres occur. Effusive basalts and rhyolites occupy small areas. In the Little Belt Mountains (Weed 1899a; 1899b; Weed and Pirsson, 1899) there are

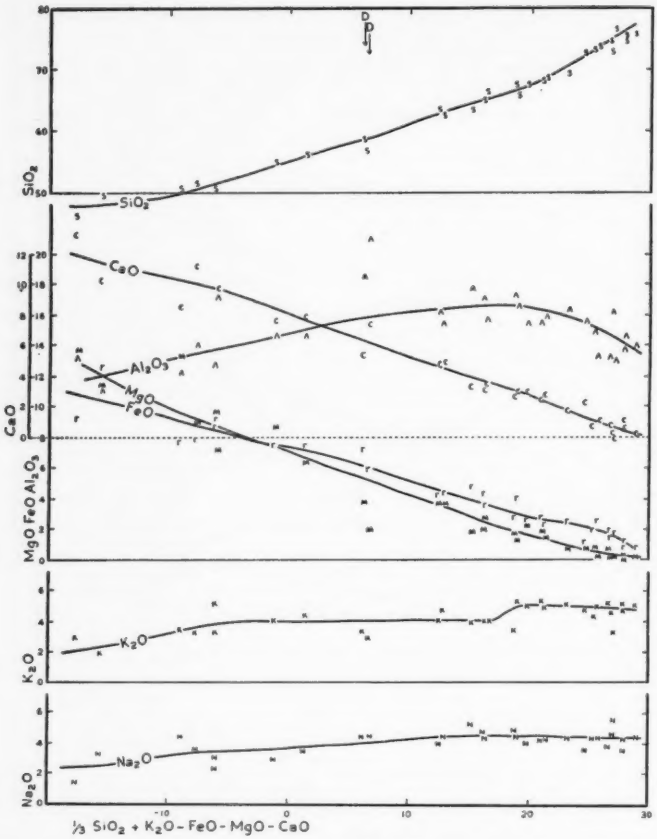

FIGURE 8.—*Variation diagram of rocks of the Little Belt subprovince*
D = Diorites with hypersthene.

diorites, granites, syenites, a single body of analcime (or nepheline) syenite, lamprophyres, and the differentiated stock of Yogo Peak, which has a plagioclase shonkinite near its border, followed toward the center by mafic monzonite, quartz syenite, and granite. The contacts are gradational. In the rocks of both mountain groups, plagioclase is abundant

in the shonkinites, the syenites, the granites, and most of the other rocks. Hornblende is abundant in many of the rocks including the shonkinite, and feldspathoids are present in a few. Hypersthene is abundant in the two diorite stocks, and these rocks do not fall on the variation curve for the subprovince and they probably belong to a different magmatic series. A variation diagram for the subprovince of the Castle and Little Belt Mountains is shown in Figure 8, and a triangular diagram for the norms in Figure 18. These rocks are much poorer in Na_2O and somewhat richer in SiO_2 than the alkalic rocks of the Crazy Mountains but are otherwise similar.

HIGHWOOD MOUNTAINS

The igneous rocks of the Highwood Mountains have been described in detail by Weed and Pirsson (1895c) and Pirsson (1905a). Hurlbut and Griggs (1939) have recently described the laccoliths, Buie (manuscript) the dikes, Burgess (manuscript) the stocks, and Larsen (manuscript) the extrusive rocks. This province has been studied more recently and in greater detail than any of the other Montana subprovinces. The writer has freely used the data of his colleagues.

The igneous rocks of the Highwood Mountains area probably belong to four subprovinces. The quartz latites that formed the older volcano are probably the oldest. The shonkinites and mafic phonolites are probably the youngest. The age relations to the other subprovinces of the syenite-monzonite stock of Highwood Peak and of the alnoite chonolith are not known.

The quartz latite formed a large volcano that was largely removed by erosion before the shonkinite-phonolite series were erupted. The analysis of the quartz latite falls near the variation diagram for the Little Belt rocks except that it is 3 per cent high in silica and 1 per cent low in soda.

The stock of Highwood Peak is made up of syenite, monzonite, and granite pegmatite. These rocks appear to be somewhat related to those of the alkalic series of the Crazy Mountains.

A group of igneous rocks that are chiefly rich in mafic minerals and are characterized by orthoclase, nepheline, diopside, biotite, and olivine with or without analcime, leucite, pseudoleucite, and sodalite, but that contain no quartz or hypersthene and rarely contain plagioclase or amphibole, make up the younger extrusive rocks of the mountains and nearly all the stocks, laccoliths, sills, and dikes. The intrusive bodies range from shonkinite to nepheline syenite and syenite; most of the extrusive rocks have the composition of shonkinite and are called mafic phonolite. The petrographic province to which this group belongs will be called the Highwood subprovince.

Thirty-two analyses of rocks from this subprovince are available and are plotted on a variation diagram (Fig. 9). They are plotted on the triangular norm diagram in Figure 19. As shown in Figure 10, they

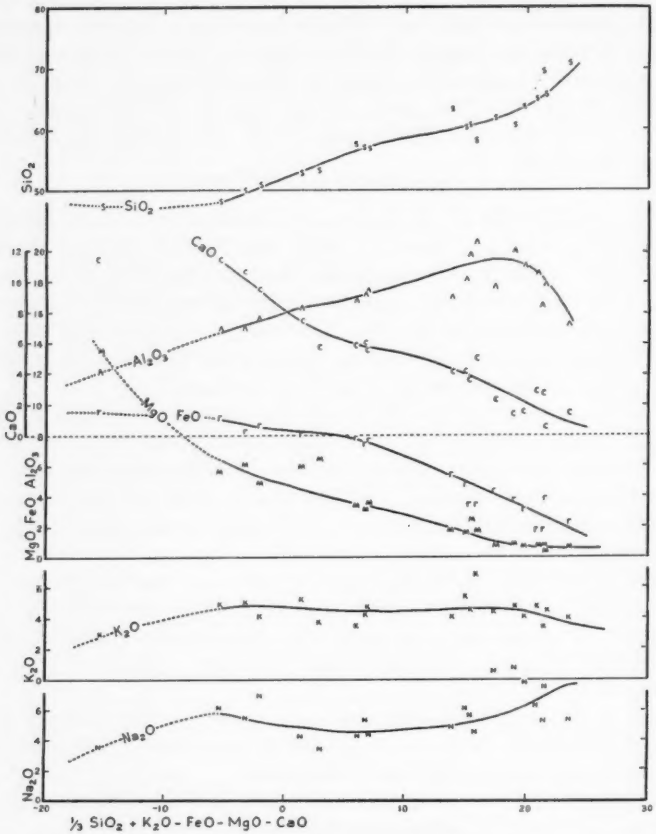

FIGURE 9.—*Variation diagram of rocks of the Highwood subprovince*
H = Subprovince of Highwood Peak.
L = Quartz latite subprovince.

are much poorer in SiO_2 and richer in Al_2O_3 and K_2O than the rocks of any of the other subprovinces which have been studied in detail. Some of the rocks of the Bearpaw Mountains appear to be richer in K_2O.

The small intrusive of Haystack Butte, near Geraldine, is made up of alnoite and monticellite peridotite, which are extremely rich in CaO.

BEARPAW MOUNTAINS

The igneous rocks of the Bearpaw Mountains are distributed over an area about 40 miles across. As a result of a trip through the mountains, Weed and Pirsson (1896a) published a brief description of the area, including seven analyses. Mr. William T. Pecora is now making a detailed study of the mountains and he has kindly furnished many of the data that follow. Further work may require some revision of the following statements.

It is probable that in this area several successive intrusions of the deep-seated magma took place, and each of these differentiated to furnish a group of related rocks. There are, therefore, several subprovinces in the area. In some respects the Bearpaw igneous rocks resemble those of the Highwood Mountains, but the former area contains a greater variety, including some richer in K_2O than any from the Highwood Mountain, plagioclase-bearing types, and more widespread, early quartz-bearing lime-alkalic varieties.

The following description was kindly furnished by Mr. William T. Pecora.

The Bearpaw Mountains and surrounding hills occupy an area of about 800 square miles of which about one half is underlain by extrusive rocks, one fourth by granular intrusive rocks, and the remainder by disturbed upper Cretaceous sedimentary rocks. The earliest Tertiary igneous activity in the area formed dikes, sills, and small plugs of quartz monzonites, syenites, and granite.

In the west-central part of the mountains typical basalt unconformably overlies the sediments and the quartz-bearing intrusive bodies; and this is in turn overlain unconformably(?) by latite. All the rocks just described are overlain unconformably by potash-rich extrusive rocks. In the northwestern part of the mountains these potash-rich lavas are separated from the underlying sediments by a thin layer of latite breccia which contains fragments of pyroxenite, garnetiferous peridotite, gabbro, and coarse syenite.

The potash-rich volcanic rocks in the mountains appear to be of two groups, both widespread. The lower group is generally badly altered and contains much natrolite, whereas the upper group is a hard, resistant, mafic leucite phonolite. Dikes corresponding to these lavas are present in and around the Rocky Boy stock.

Intrusive feldspathoid-bearing rocks are almost entirely concentrated in the western part of the mountains and are present as dikes, sills, laccoliths, and two small and one large composite stocks. Their relations are best observed in the large Rocky Boy composite stock, about 14 square miles in area, located at the headwaters of Big Sandy, Beaver, Box

Elder, and Eagle creeks in the southern part of the Rocky Boy's Indian Reservation. Some of these relations are also seen in the small Beaver Creek composite stock partially described by Weed and Pirsson (1896a). There appear to have been two periods of intrusion of the feldspathoid-bearing rocks separated by the intrusion of a hard, fine-grained olivine-hornblende monzonite which itself has a strong alkalic affinity.

The earlier of the feldspathoid-bearing intrusions is composed of peridotites, pyroxenites, shonkinites, and syenites, all of which are generally coarse-grained but lacking in pegmatitic or mineral deposits. All the shonkinites and syenites have at least some nepheline or pseudoleucite, except a few which have coarse plagioclase (labradorite). The plagioclase shonkinite and associated plagioclase syenite present on the western and southwestern fringe of the mountains are interpreted as falling into this early alkalic series. The differentiated laccolith east of Box Elder is much like the differentiated laccoliths of the Highwood Mountains except that the shonkinite and syenite near Box Elder contain plagioclase.

The olivine-hornblende monzonite which interrupts the sequence of the feldspathoid-bearing series is rather widespread in the western part of the mountains. These monzonites range from very mafic to leucocratic hornblende-rich rocks. Two analyses of this monzonite do not show very high potash.

The second group of feldspathoid-bearing intrusions, containing high potash and volatiles, are localized within the Rocky Boy composite stock. Unusual pegmatites and veins characterized by minerals high in titanium, strontium, and zirconium are genetically associated with these rocks which are typical nepheline syenite, nepheline-sodalite syenite, and syenite.

The last phase of the intrusive alkalic magma is indicated by numerous dikes of coarse pseudoleucite tinguaite porphyry within and around the composite stocks. These dikes are younger than any of the intrusions mentioned above.

Numerous dikes of mafic leucite phonolite equivalent to the resistant leucite phonolites are intrusive into the older volcanic rocks, the early intrusive of feldspathoid-bearing rocks, and the olivine-hornblende monzonite. Relations have not been observed between these dikes and the younger intrusive feldspathoid-bearing series. The altered basal, potash-rich lavas and tuffs unconformably overlie stocks of the olivine-hornblende monzonite or their associated structures.

It appears that a long time interval separated the early feldspathoid-bearing intrusions from the earlier basaltic rocks. Much shorter time intervals probably separated the olivine-hornblende monzonite intrusions

from the early and late intrusive feldspathoid-bearing series. Finally, the eruptions of the potash-rich lavas were preceded by a well-recognized erosional interval.

There is a conspicuous difference with respect to the alteration of the nepheline between the rocks of the Highwood Mountains and those of the Bearpaw Mountains. In the Highwood Mountains nepheline was originally present in most of the rocks, part of it in the pseudoleucite. With rare exceptions it is now altered, chiefly to a cloudy amorphous material that has about the composition of nepheline with added water. In the Bearpaw Mountains, on the other hand, fresh nepheline is common at least in the dikes and stocks which have been examined so far (Mr. Pecora, personal communication), both in the pseudoleucite and in other associations. The two mountain ranges are much alike in respect to relief and rate of erosion. If the alteration of the nepheline was paulopost, the difference in alteration in the two ranges may be due to a difference in the quantity or character of the mineralizers.

AREA WEST OF CASCADE

On both sides of the Canyon of the Missouri River, on the northern end of the Big Belt Mountains southwest of the town of Cascade, there is a mountain range underlain by lavas and breccias. Numerous dikes, sills, and stocks cut the lavas and the underlying Cretaceous sedimentary rocks. This igneous complex is probably larger than that of the Highwood Mountains. A study of this area is being made by Mr. J. B. Lyons, who has furnished the following statement.

The volcanics are for the most part clastic beds made up of fragments of dark rocks with large phenocrysts of olivine and pyroxene. Megascopically they resemble the rocks of the Highwood Mountains. Some are lighter-colored. The microscope shows that nearly all are much altered to zeolites and similar products. They are rich in calcic plagioclase, contain some interstitial orthoclase, and are trachydolerites. Some contain phenocrysts of clear analcime. Interbedded with the trachydolerite breccia south of the Missouri River and at an elevation of about 2000 feet above the bed of the river are flows of a mafic analcime phonolite much like those of the Highwood Mountains. The rocks of the Big Belt Mountains closely resemble those of the Absaroka subprovince of the Yellowstone Park area. They appear to be even richer in calcic plagioclase and may be rich in soda.

SWEET GRASS HILLS

The rocks of the Sweet Grass Hills, in the extreme northern part of Montana, have been described by Weed and Pirsson (1895a) and by

Kemp and Billingsley (1921). There are three large stocks in the Hills and a number of smaller intrusive bodies. The western stock is made up of quartz diorite porphyry with some orthoclase and with hornblende as the characteristic mafic mineral; the middle stock is augite diorite porphyry with some hornblende; and the eastern stock is a quartz syenite porphyry with plagioclase and aegirine. The rocks of the dikes and sills are tinguaites, minettes, and other lamprophyres. These rocks belong to a plagioclase-hornblende series somewhat like that of the Little Belt Mountains. Some of the dikes may belong to the Highwood type.

INTRUSIVE ROCKS OF THE MOUNTAINS NEAR LEWISTON

The mountains near Lewiston will be described in order from south to north. The Big Snowy Mountains have been described by Reeves (1927; 1931), who states that they represent a dome in the sediments and are probably underlain by a laccolith. A few small dikes of an altered brick-red rock in the eastern part and a few small breccia pipes of rhyolite are present. About 40 miles east of Lewiston, and 11 miles west of Winnett, is a sill of alnoite (Reeves, 1927, p. 54-55; Ross, 1926).

In the Judith Mountains Weed and Pirsson (1897) found several complex laccoliths, the largest of which is 6 miles across. The chief rock of these laccoliths is granite porphyry. There is also syenite, diorite, and a syenite with plagioclase (monzonite, near the yogoite of Yogo Peak, in the Little Belt Mountains). A later group of phonolitic rocks includes aegirite syenite porphyry and tinguaite. The laccoliths appear to be made up of rocks much like those of the Little Belt Mountains. A tinguaite was analyzed and is of the Highwood Mountains type.

In the South Moccasin Mountains Palmer (1925) described a laccolith of rhyolite porphyry 6 miles across.

In the North Moccasin Mountains Blixt (1933) found a laccolith of quartz syenite 2½ miles across. Lindgren (1893) described porphyrite and quartz porphyrite from the Moccasin Mountains.

In the Little Rocky Mountains Weed and Pirsson (1896b) and Emmons (1908b) have described a laccolith 6 miles across made up chiefly of a plagioclase-bearing granite porphyry with some syenite, granite-diorite, and tinguaite. The analyzed granite is lower in SiO_2 and iron and higher in Al_2O_3 than most of the granites of the Central Montana province. It has 3.47 per cent corundum in the norm, which casts doubt on the freshness of the rock or the accuracy of the analysis.

CHEMISTRY

Chemical analyses of a sufficient variety of rocks to serve for a satisfactory study and for the construction of variation diagrams are avail-

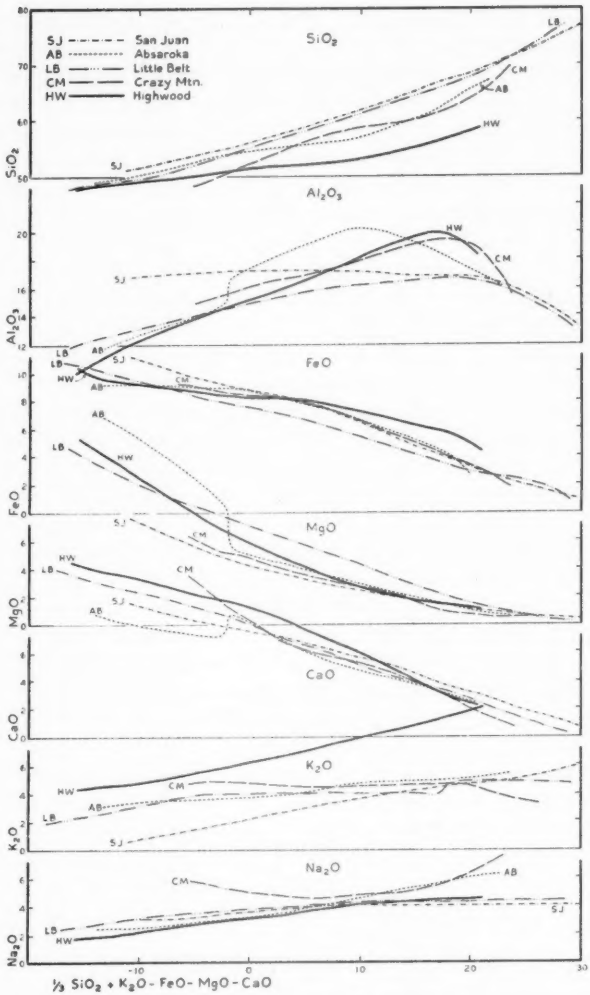

FIGURE 10.—*Variation curves for rocks of six subprovinces*
Central Montana petrographic province and the San Juan Mountains, Colorado, province.

able for the three subprovinces of the Yellowstone area, for the Crazy, Castle, and Little Belt Mountains, and for the potash-rich subprovince of the Highwood Mountains. Variation diagrams for these subprovinces

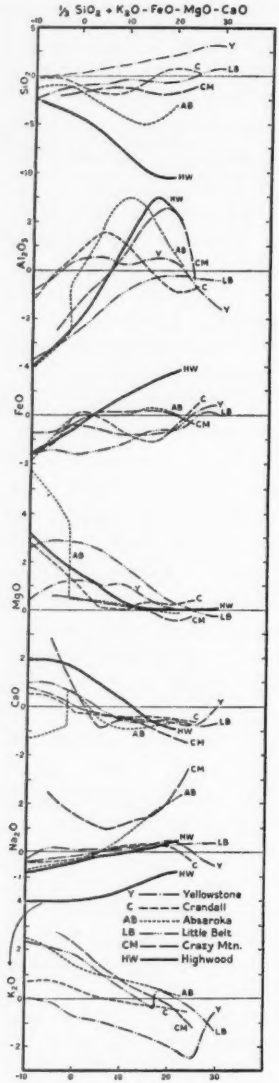

FIGURE 11.—*Comparison of the petrographic subprovinces of Central Montana and that of the San Juan Mountains*

are shown in Figures 2-4 and 6-9, and superimposed variation curves for these subprovinces and for the lime-alkalic province of the San Juan, Colorado, lavas are shown in Figures 5 and 10. In Figure 11 a comparison of the various subprovinces is made. On this diagram the horizontal scale is the same as that of the variation diagrams, and the vertical scale is four times as large. A separate curve is drawn for each oxide for each subprovince. The vertical scale is the amount the curves of each oxide differ from those of the San Juan province. The San Juan province is, therefore, taken as zero.

Most of the subprovinces are low in silica throughout. The Yellowstone rocks are like those of the San Juan at the mafic end and are 3 per cent high at the felsic end. The Highwood rocks are about 2½ per cent low at the mafic end and 10 per cent low at the felsic end. The Crandall rocks are near those of the San Juan at both ends and are low in the intermediate range. The Little Belt rocks are 2½ per cent low at the mafic end and are near those of the San Juan at the felsic end. The Absaroka and Crazy Mountains rocks are low throughout.

The rocks of the Yellowstone and Crandall subprovinces contain about as much Al_2O_3 as do those of the San Juan Mountains except that the Crandall rocks are nearly 2 per cent high in the intermediate (andesitic) rocks. In most of the subprovinces the Al_2O_3 difference is low at the mafic end, rises to a maximum in the intermediate rocks, and falls to nearly 0 at the felsic end. This maximum for Al_2O_3 in the intermediate rocks is especially prominent in the alkalic subprovinces and is due to the presence of syenites, monzonites, nepheline syenites, and similar rocks.

The MgO for all but the Yellowstone subprovince is high at the mafic end and nearly the same as for the San Juan province at the felsic end. The Yellowstone subprovince has the same MgO content as the San Juan province at the two ends but is high in the intermediate rocks. The MgO of the Absaroka subprovince is very high at the mafic end and shows a sharp decline at position —1½ (*see* Figure 5), beyond which it is nearly like that of the San Juan province. At this point (—1½) Al_2O_3 and CaO show less sharp breaks than does MgO. The other curves have no perceptible breaks.

For CaO the Absaroka rocks are somewhat low; the others are high at the mafic end and about the same or a little low at the felsic end.

For FeO most of the curves are low at the mafic end and at about 0 at the felsic end. The Highwood curve has a rather uniform slope and is about 2 per cent low in the mafic rocks and 2 per cent high in the felsic rocks.

For most of the subprovinces Na_2O is in about the same amount as in the San Juan province. It tends to be low in the mafic rocks and high

in the salic rocks. For the Crazy Mountains rocks Na_2O is about 2 per cent high over most of the range. For the Absaroka rocks it is 0.6 per cent low at the mafic end and 2 per cent high at the felsic end.

K_2O is from 4 to 5 per cent high in the Highwood rocks. It is from 2 to 3 per cent high at the mafic end of the Little Belt, Crazy Mountains, and Absaroka subprovinces but is about like the San Juan or low at the felsic end. It is low except at the two ends for the rocks of the Yellowstone subprovince. It is a little high at the mafic end and a little low at the felsic end in the Crandall subprovince.

The rocks of the Yellowstone subprovince are much like those of the San Juan Mountains but they are high in SiO_2 at the felsic end, low in Al_2O_3 at the extreme salic end (in the rhyolites), and low in K_2O and FeO and high in MgO in the intermediate rocks.

The rocks of the Crandall subprovince are nearer to those of the San Juan province than any of the others. At the mafic end they are high in MgO and low in FeO.

The Absaroka, Little Belt, and Crazy Mountains subprovinces are similar but have some marked differences. The SiO_2 is low in all three subprovinces except that it is near zero at the felsic end for the Little Belt subprovince. For all three subprovinces Al_2O_3 is very low at the mafic end and near zero at the felsic end. MgO is very high at the mafic end except for the Crazy Mountains rocks and at about zero at the salic end. In the Crazy Mountains and Little Belt subprovinces CaO is high at the mafic end and a little low at the felsic end. In the Absaroka subprovince CaO is low except near the break in the curve. FeO is low at the mafic end. In the Crazy Mountains subprovince Na_2O is very high. In the Absaroka rocks it is low at the mafic end and very high at the felsic end. In the Little Belt rocks it is at zero at the mafic end and at 0.8 at the felsic end. For the three subprovinces K_2O is much the same and is high at the mafic end and near zero at the felsic end.

The Highwood subprovince has very low SiO_2 and high Al_2O_3 at the felsic end as the felsic rocks are rich in feldspathoids, and no quartz-bearing rocks are represented. MgO and CaO follow the usual course and approach zero in actual amount at the salic end. FeO, however, is low in the mafic and high in the felsic rocks. Na_2O is nearly the same as in the San Juan rocks, but K_2O is very high.

The data available for the rocks of the Bearpaw Mountains indicate that the late phonolite extrusive rocks and the Rocky Boy stock are near the rocks of the Highwood Mountains but even higher in potash. Some of the groups of rocks in the Bearpaw Mountains are nearer those of the Little Belt or Crazy Mountains. The trachydolerites at the

northern end of the Big Belt Mountains are much like those of the Absaroka subprovince. However, some of the rocks of this mountain group are not very different from the potash-rich rocks of the Highwood Mountains.

A comparison of the curves for MgO and FeO shows that, in general, where MgO is high FeO is low. This holds fairly well for individual analyses. High K_2O tends to be associated with high CaO and MgO and low FeO.

The oxides, except SiO_2 and Na_2O for most of the subprovinces and K_2O, Al_2O_3, and FeO for the Highwood subprovince, approach a common value at the felsic end. This is especially true of the CaO, FeO, and MgO which are approaching zero at this end, and for K_2O which is near 5 per cent.

TRIANGULAR DIAGRAMS FOR THE NORMS

Triangular diagrams of the norms after Larsen (1938) of the San Juan province of Colorado, and of the Yellowstone, Crandall, Absaroka,

TABLE 2.—*Characteristic features of the triangular diagrams of the norms*
Petrographic subprovinces of Central Montana and of the San Juan province of Colorado.

Subprovince	Line		Ab at			Or at		Max. An Content	SiO_2 at		Ratio Fem/An in feldspar
	Length	Slope*	An_5	An_{30}	An_{40}	An_5	An_{30}		fm_5	fm_{30}	
Yellowstone.....	20+	+55°	55	50	45	40	20	45	30 Low	10 at 30	½
Crandall.....	25	+25°	55	50	40	40	20	40	23	0	1
San Juan.....	25	+20°	50	45	43	45	37	50	30	10	1
Little Belt†.....	20	+ 5°	53	48	.....	42	.....	25	20	0	1
Absaroka†.....	35	+ 5°	60	52	38	35	36	40	10	−5	1
Crazy Mountains†	33	−10°	60	50	.....	30		22	2	−3	near 1
Little Belt steep..	40	−35°	40		.....	40		23	low		2+
Absaroka steep...	40	−35°	35		.....	40		25	low		2
Crazy Mts. steep..	48	−45°	65	55	.....	35		17	very low		very large
Highwood.....	45	−20°	45	25	.....	50		20	−6	−15	1 to 10

* Slopes down to right are +, down to left are −.
† Exclusive of the few with very steep slope.

Crazy Mountains, Little Belt, and Highwood subprovinces of Montana and Wyoming are shown in Figures 12-19. In these diagrams for each analyzed rock the normative feldspar is shown by a dot, the proportions of normative feldspar, femic and quartz or silica deficiency are represented by a circle, and dot and circle are joined by a line.

The characteristic features of these diagrams and of the norms for the various provinces are shown in Table 2. All the data are estimates.

The meaning of these data is somewhat complex. In general, long lines indicate low silica content for the composition of the feldspar. A steep slope to the left (—) indicates high content of femic for the anorthite con-

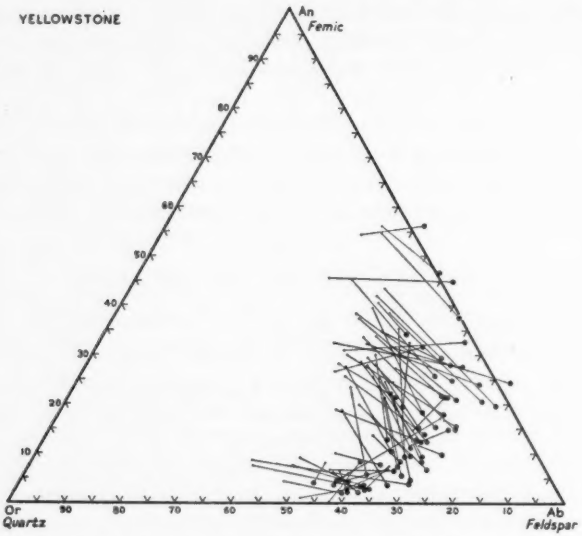

FIGURE 12.—*Triangular diagram of the norms for the Yellowstone subprovince*

Dots = Composition of normative feldspar.
Circles = Feldspar-femic-quartz or deficiency in SiO_2.

tent of the feldspar. A slope to the right (+) indicates low content of femic for the anorthite content of the feldspar.

BREAKS IN THE VARIATION CURVES

Of variation diagrams for about 30 petrographic provinces that the writer has drawn, that of the Absaroka subprovince, shown in Figures 4 and 10, has the most sharp breaks in the curves. At the position —2 the MgO abruptly increases about 3 per cent, the Al_2O_3 decreases about an equal amount, and the CaO decreases about 1½ per cent. There is a flattening of the curve for iron at this position. The other curves show no obvious breaks, although breaks are probably present. This break takes place at about the point to the right of which rocks and analyses are numerous and to the left of which they are scattered. In most variation diagrams of petrographic provinces with a wide range in rock compo-

sition, a part of the diagram has the analyses closely spaced, with a tendency for an unusually close spacing at the ends of this range. Beyond this range of close spacing, especially at the mafic end, analyses are few and do not fall near a regular curve. The mafic end of the close

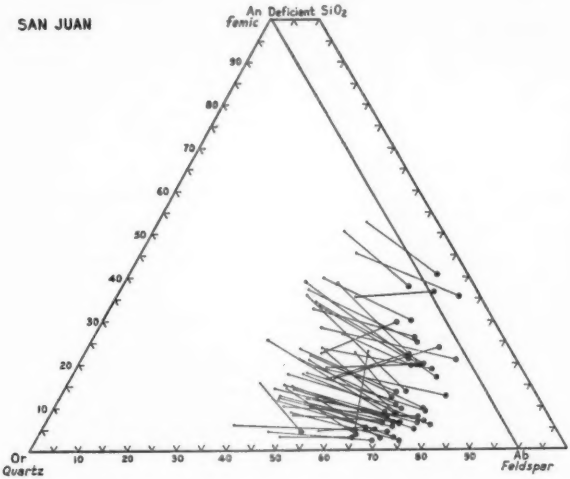

FIGURE 13.—*Triangular diagram of the norms for the San Juan, Colorado, province*

Dots = Composition of normative feldspar.
Circles = Feldspar-femic-quartz or deficiency in SiO_2.

spacing is believed commonly to represent approximately the parent magma of the province.

For the Absaroka subprovince the break is clearly shown in the triangular diagram of the norms (Fig. 15), as the rocks to the left of the break are represented by the long lines that slope steeply to the left. These rocks are leucite basalts, orthoclase basalts, orthoclase gabbro, and gabbro.

All the rocks classed as absarokites by Iddings and others fall to the left of the break in the curves, and only one rock falling to the left of the break was not classed as absarokite. This latter is an olivine gabbro from the White Mountain stock in the Sunlight area. It does not fall on the variation diagram. The monzo-gabbro of the ring-dike of Sulphur Creek Basin falls on the very steep slope of the MgO curve.

On the variation diagram for the alkalic rocks of the Crazy Mountains (Fig. 7) there is a rapid change in the CaO at about the position 0.

A possible change in Na_2O takes place at this position. There are too few analyses to show the details of the change. On the triangular diagram for these rocks (Fig. 17) a change in the character of the lines takes place at this point, and those to the right (analysis 1, 2, 3, and 4)

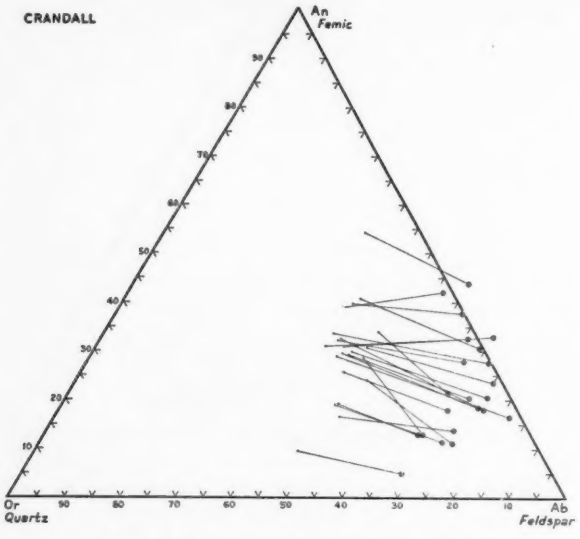

FIGURE 14.—*Triangular diagram of the norms for the Crandall subprovince*

Dots = Composition of normative feldspar.
Circles = Feldspar-femic-quartz or deficiency in SiO_2.

are long and dip steeply to the left. Analysis 5, which is along the steep part of the CaO curve, is intermediate in length and slope. The four rocks represented by points to the left of the break are shonkinites (theralites of Wolff). Number 5, along the break, is from a dike on Sweet Grass Creek.

The variation diagram for the Little Belt subprovince shows no break corresponding to that of the Absaroka subprovince. However, the triangular diagram has the long steep lines for all the fresh rocks for which norms have been calculated and which are to the left of position +5 on the variation diagrams. On the triangular diagram analysis 9 is intermediate between the short, nearly horizontal lines and the long steep lines. This represents the syenite from the Yogo Peak differentiated

stock. The yogoite (monzonite) and shonkinite from this stock are represented by lines 12 and 14 respectively. The other rocks represented by the long lines are minette, monchiquite, and a basalt flow from the Castle Mountains.

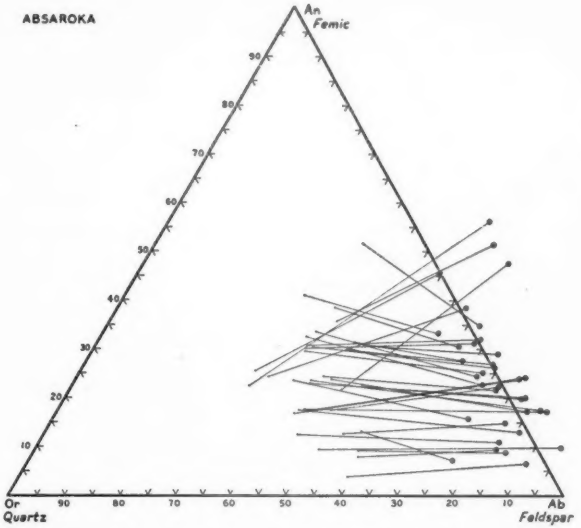

FIGURE 15.—*Triangular diagram of the norms for the Absaroka subprovince*

Dots = Composition of normative feldspar.
Circles = Feldspar-femic-quartz or deficiency in SiO_2.

On the Little Belt diagram (Fig. 18) there appears to be a gradual transition from the short flat lines of the triangular diagram to the long steep lines, shown especially by the rocks of the Yogo Peak differentiated stock. The rocks represented by long, steep lines have a low anorthite content in the feldspar, are rich in femic minerals, and poor in SiO_2. They are much like the rocks of the Highwood province but are richer in femic minerals. On the whole, they do not fall as close to variation curves as do the other rocks.

These rather abrupt changes in the character of the rocks at the mafic side of the variation and triangular diagrams might be explained in a number of ways. Only those believed to be more probable will be considered. The breaks might be explained as due to the addition of settled crystals to a magma on the right of the breaks. The effect of adding crys-

tals of augite and olivine to a magma is well shown in the laccoliths of the Highwood Mountains. No break in the curves results even where a high percentage of such crystals is added. This is to be expected if the differentiation was caused by crystal settling; the residual liquid would

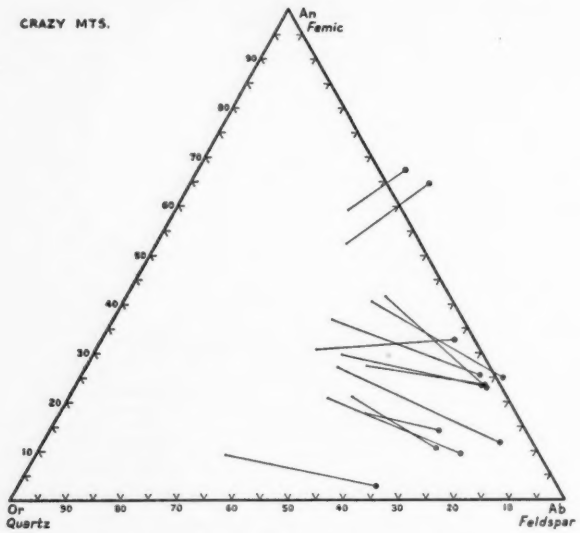

FIGURE 16.—*Triangular diagram of the norms for the southern (lime-alkalic) subprovince of the Crazy Mountains*

Dots = Composition of normative feldspar.
Circles = Feldspar-femic-quartz or deficiency in SiO_2.

have lost the same material as the crystal mush gained, and only where there was a concentration of one of the crystals in one part of the crystal-enriched magma would a break result.

The addition of about 15 per cent of iron-poor hypersthene to the Absaroka magma represented by a point just to the right of the break would cause the break. A serious objection to this explanation is the fact that the residual liquid after the loss of hypersthene would not fall on the variation curve unless the original liquid was along the vertical part of the MgO curve. This would not explain the break, as the rapid crystallization of hypersthene along the vertical part of the curves with little crystallizing on either side would also be required.

Another explanation is by crystal settling. The curves would be about as found if from the absarokite magma at the extreme left olivine,

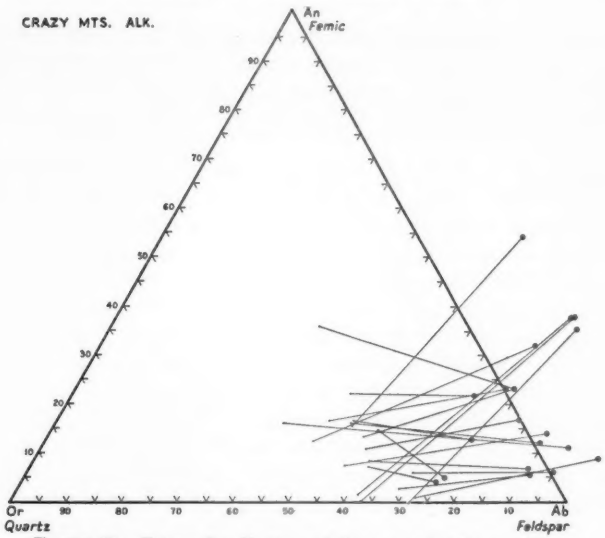

FIGURE 17.—*Triangular diagram of the norms for the northern (alkalic) subprovince of the Crazy Mountains*

Dots = Composition of normative feldspar.
Circles = Feldspar-femic-quartz or deficiency in SiO_2.

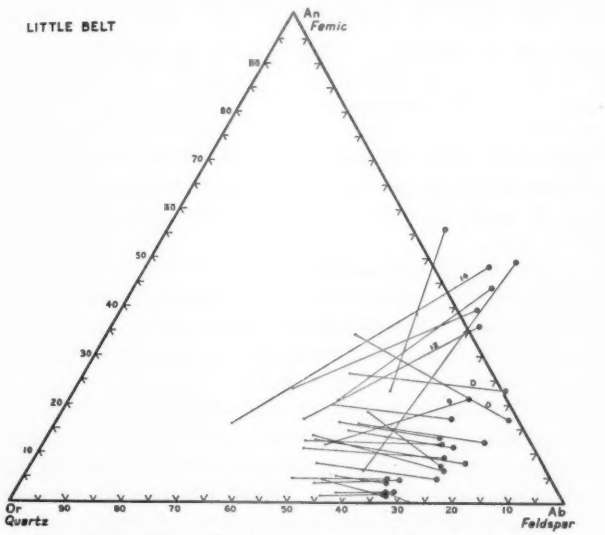

FIGURE 18.—*Triangular diagram of the norms for the Little Belt subprovince*

Dots = Composition of normative feldspar.
Circles = Feldspar-femic-quartz or deficiency in SiO_2.

plagioclase, and augite crystallized and settled along the part of the curve to the left of the break, and at the break hypersthene began to crystallize in large amount, continued along the break, and ceased at the lower part of the break, with olivine and hypersthene continuing beyond

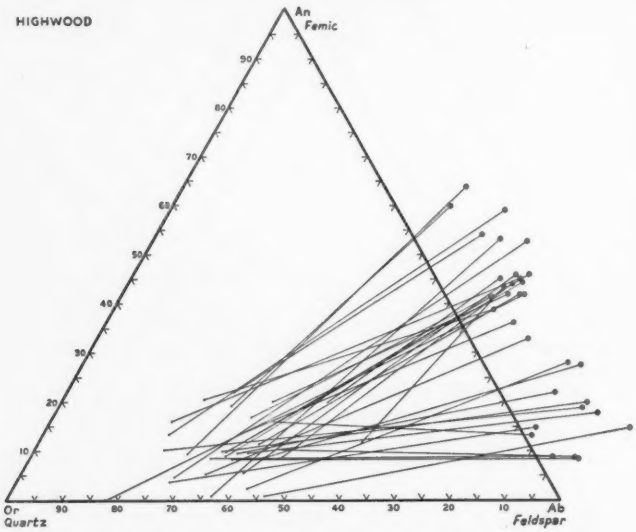

FIGURE 19.—*Triangular diagram of the norms for the Highwood subprovince*

Dots = Composition of normative feldspar.
Circles = Feldspar-femic-quartz or deficiency in SiO_2.

to the right. Since the phenocrysts in the Absaroka rocks are olivine and augite to the left of the break and olivine, augite, and plagioclase to the right, it would be more reasonable for olivine with some augite crystals to settle.

In the Absaroka subprovince the analyses to the left of the break may represent a somewhat different but closely related primary magma from those to the right. Objections to this are the facts that several analyses fall along the steep line of the break, and one curve ends where the other begins.

On the variation diagram for the Little Belt subprovince there is an abrupt decrease in the K_2O and a less clear increase in Na_2O at position 19. Rocks from both the Little Belt and Castle Mountains are represented on both sides of the break. On the triangular diagram for the Little Belt subprovince a group of rocks near the bottom of the diagram

are high in quartz. These are from the Castle Mountains; those near-by and above are from the Little Belt Mountains.

MAGMATIC DIFFERENTIATION

ORIGIN OF THE PRIMARY MAGMAS OF THE SUBPROVINCES

The preceding descriptions show that the Central Montana petrographic province is made up of several subprovinces. Some subprovinces represent the Tertiary igneous rocks of some area or mountain range, but others include only the rocks erupted during a limited time interval in a mountain range. In the Highwood Mountains four subprovinces have been recognized. The rocks of several widely separated areas may be very similar. Where the age relations of the subprovinces are known, the earliest are commonly ordinary lime-alkalic rocks, and later subprovinces become progressively nearer the extreme potash-rich alkalic rocks. The Yellowstone National Park area is an exception to this. The subprovinces probably represent eruptions during a time interval rather than within a limited area. In the subprovinces for which data are available, the earliest emplaced, the original magma of the differentiated laccoliths, and in many of the areas the commonest rocks were of a mafic magma ranging from gabbro to shonkinite.

Estimates of composition of the original magmas of the San Juan lavas of Colorado and of the subprovinces of the Central Montana area have been made from the variation diagrams taking into account the position near the mafic ends beyond which few rocks are known or beyond which the rocks do not fit a regular variation curve. These data are given in Table 3. In this table the chief oxides are listed as they are used in the variation diagram; total iron plus MnO is calculated to FeO, BaO and SrO are added to CaO, and the sum of the listed oxides is adjusted to 100 per cent. Below the table some of the minor constituents are listed. The provinces are listed in order of increasing K_2O content, which is approximately the order of increasing difference from the San Juan magma.

The most conspicuous differences between the San Juan magma and those of the Montana subprovinces are the high K_2O and low Al_2O_3 in most of the latter. In the magmas of the Montana province FeO is consistently low. MgO tends to be high in the magmas of the Crandall, Little Belt, and Highwood subprovinces. CaO and Na_2O are high in the Crazy Mountains and Highwood magmas. BaO and SrO are very high in all the magmas of the Montana province and they tend to increase with the K_2O. MnO is about the same in all the magmas. TiO_2 is low and P_2O_5 somewhat high in the magmas of Montana province, although in many femic alkalic rocks the TiO_2 and P_2O_5 are very high.

In order to study the origin of the magmas of the subprovinces of the petrographic province of Central Montana, the author has calculated (1) the composition of the smallest amounts of material that must be added to the San Juan magma to yield the postulated parent magmas of the

TABLE 3.—*Estimated composition of the original magmas*
Of the subprovinces of the Central Montana petrographic province and of the San Juan lavas of Colorado.

	Yellowstone	Crandall	Absaroka		Little Belt	Crazy	Highwood	San Juan	Average Plateau Basalt
			at −10	at −1.5					
SiO_2	56.0	52.8	50.6	53.8	53.8	48.8	48.7	53.8	51.2
Al_2O_3	17.7	15.5	12.9	16.6	14.5	15.0	12.7	17.2	14.7
FeO	8.1	8.4	9.3	8.9	8.0	9.1	9.0	9.8	13.9
MgO	5.6	9.0	13.3	5.4	7.8	5.2	9.2	5.8	7.0
CaO	7.6	9.2	8.1	8.8	8.5	11.1	11.0	8.4	9.8
Na_2O	3.7	2.8	2.5	3.0	3.6	5.9	4.4	3.4	2.7
K_2O	1.6	2.0	3.3	3.6	4.0	4.8	5.2	1.6	0.7
BaO	.1	.....	.....	.2	.4	.4	.5	.06	.....
SrO	tr.	.....	.....	.1	.1	.1	.2	.03	.....
MnO	.12	.15	.....	.1	.1	.1	.1	.15	.17
TiO_2	.7	.....	.....	.8	.8	.9	.6	1.40	2.17
P_2O_5	.3	.4	.....	.6	.7	.9	.8	.40	.33

various subprovinces, and (2) the composition of the smallest amounts of material that must be subtracted from the San Juan magma to yield the postulated parent magmas of the subprovinces. These data are given in Tables 4 and 5. The San Juan province was selected for comparison, since it represents a large body of Tertiary volcanic rocks which vary widely in composition and which have been carefully studied by Whitman Cross and the author. They are in the Rocky Mountains and are believed to be genetically related to the rocks of the Montana area.

The materials that must be added to the San Juan magma (Table 4) to form the magmas of the subprovinces of the Montana province are not large, reaching 27 per cent for the Highwood magma. For the different subprovinces this material is rather variable but for all it is high in alkalies and CaO and low to very low in Al_2O_3. For all four subprovinces the Al_2O_3 is in far smaller amount molecularly than the alkalies. Materials of these compositions could be supplied by the assimilation of potash salt deposits but not by the assimilation of or selective reaction on any ordinary rock or by the mixing of any magma with a basaltic magma. Assimilation or mixing of two magmas, therefore, is not a probable origin for the parent magmas of the subprovinces. Such material might be carried by gas transfer (Fenner, 1926, p. 715-744), but gas transfer on a large scale seems improbable at great depth.

As shown in Table 5, from 55 to 69 per cent of material must be subtracted from the San Juan magma to yield the magmas of the subprovinces of the Montana area. The norms of these materials are chiefly

TABLE 4.—*Composition of the smallest amounts of material that must be added to San Juan basalts to yield postulated parent magmas of subprovinces of petrographic province of Central Montana*

	Absaroka	Little Belt	Crazy	Highwood
SiO_2	54.0	53.8	15.5	34.8
Al_2O_3	10.9	2.6	0	0
FeO	1.0	0	5.2	6.6
MgO	1.5	16.1	1.6	18.2
CaO	12.0	9.1	29.5	18.6
Na_2O	0	4.3	22.5	7.1
K_2O	21.2	14.8	26.0	15.2
Amount	10	18	13	27

TABLE 5.—*Composition of the smallest amounts of material that can be subtracted from San Juan basalts to yield postulated parent magmas of alkalic subprovinces of petrographic province of Central Montana*

	Absaroka	Little Belt	Crazy	Highwood	
SiO_2	53.8	53.8	56.5	55.8	
Al_2O_3	17.7	19.0	18.5	19.2	
FeO	10.6	11.0	10.1	10.2	
MgO	6.0	4.5	6.1	4.2	
CaO	8.0	8.3	7.0	7.2	
Na_2O	3.6	3.3	2.2	2.9	
K_2O	0	0	0	0	
Amount	55	60	67	69	
		Norm			Add 7% CaO to Highwood
Q	.....	2.	11.	9.	.....
cor	.....	.....	2.	1.	.....
ab	30.	28.	18.	25.	24.
an	32.	37.	35.	36.	37.
di	6.	3.	.....	.....	26.
hy	31.	30.	34.	29.	15.
Mol. ratio MgO/FeO	1.0	0.7	1.1	0.7	0.7

calcic plagioclase and hypersthene and are much the same for all the subprovinces. Some of the norms show quartz. The calculated norms consider all the iron as FeO. If some of the iron crystallized as an oxide, more quartz would result.

The average plateau basalt of Daly (1933, p. 17) falls very near the San Juan variation curves, slightly projected to the mafic side, except

that it is somewhat low in MgO and Al_2O_3 and very high in FeO (over 2 per cent). The composition of the least amounts of material that can be removed from the average plateau basalt to yield the postulated parent magmas of the various petrographic provinces discussed in this

TABLE 6.—*Composition of the smallest amounts of material that can be removed from the average plateau basalt to yield postulated magmas of various petrographic provinces*

	San Juan	Yellowstone	Crandall	Absaroka		Little Belt	Crazy	Highwood
				at −10	at −1½ Right of break			
SiO_2	49.1	47.0	50.0	51.4	50.6	50.9	51.6	51.6
Al_2O_3	12.6	12.2	14.2	15.2	14.2	14.8	14.6	15.0
FeO	17.4	18.4	16.8	15.2	15.1	15.1	14.9	14.7
MgO	8.0	8.1	5.8	5.3	7.4	6.8	7.3	6.6
CaO	10.8	11.5	10.1	10.2	10.0	10.0	9.5	9.6
Na_2O	2.1	2.9	2.7	2.8	2.6	2.5	2.1	2.4
Amount	56	56	65	79	81	83	85	87
Norms								
ne	.....	5	.....	.....	.....	.....	.....	.....
ab	18	15	23	24	22	21	18	20
an	25	20	26	29	27	30	30	30
di	24	31	20	18	19	17	14	15
hy	14	0	16	22	18	23	35	31
ol	19	29	14	8	14	10	2	4
Mol. ratio MgO/FeO	.83	.80	.62	.63	.88	.81	.88	.81

report are shown in Table 6. Since the plateau basalt is somewhat similar in composition to the postulated parent magmas and since a large amount of material must be removed, the material removed is not very different in composition from the plateau basalt itself. The high iron content of the plateau basalt makes it necessary to remove much FeO.

The material calculated to be subtracted from the plateau basalt is much greater in amount than that from the San Juan basalt and is lower in SiO_2, Al_2O_3, and MgO, higher in CaO, and much higher in FeO. Its norms have no quartz and some olivine, and less feldspar and more diopside than those derived from the San Juan basalt.

A calculation of the material that must be removed from the olivine basalt, which Kennedy (1933, p. 241) believes to be the parent magma

of the alkalic rocks, to form the parent magmas of the potash-rich subprovinces of Central Montana is about the same for all the subprovinces and is a little low in MgO for the Absaroka magma at —10. The average composition is:

		Norm	
SiO_2	48	ab	21
Al_2O_3	16½	an	34
FeO	15	di	11
MgO	9	hy	7
CaO	9½	ol	27
Na_2O	2½	Mol. ratio MgO/FeO = 1.1	

This material is not very different from that which must be removed from the plateau basalt but it is lower in SiO_2 and higher in Al_2O_3 and MgO.

The calculations made in Tables 5 and 6 and in the preceding paragraph are not very accurate but they are believed to be fair approximations and to represent the general character of the materials that must be subtracted from the San Juan basalt, the average plateau basalt, or the olivine basalt to yield the parent magmas of the various subprovinces. One source of error is in the selection of the parent magmas for the various petrographic provinces, and the magmas selected may represent magmas that have already undergone some differentiation near the surface. If a more mafic composition had been selected for the San Juan basalt, more material would have to be removed and it would be lower in SiO_2.

The presence of quartz in the minerals that must be removed from the San Juan basalt to form the parent magmas of the Montana province (*see* Table 5) introduces a serious difficulty unless we postulate that the San Juan basalt was a more femic type. The low ratio of MgO to FeO in both Tables 5 and 6 introduces another difficulty and would require that the femic minerals removed were iron rich. If some of the iron was removed as magnetite, this would reduce the iron content of the femic minerals, but it would increase the quartz in the minerals removed from the San Juan magma and would decrease the olivine in those removed from the plateau basalt. If the olivine basalt were the parent magma, these difficulties would be less.

Under surface conditions, calcic plagioclase would crystallize early from the magmas of any of the three basalts, but we expect olivine and diopside to be the early femic minerals, especially since we have abundant olivine crystallizing from the shonkinites of the Montana province. The differentiation of the basaltic magma to form the shonkinitic types is believed to have taken place at great depth, and the crystallization of

a magma under the resulting pressure would differ from the crystallization near the surface. It is probable that high pressure would favor the crystallization of hypersthene in place of olivine. One who has attempted to separate by heavy liquids the pyroxenes and olivines from an igneous rock knows that they commonly have very nearly the same specific gravities. They also have very nearly the same ratios of MgO to FeO. It is very probable that the density of hypersthene is greater than that of a mixture made up of olivine and liquid which together have the composition of hypersthene. This would probably be true if a basaltic liquid is added to both systems. Since high pressure favors the denser system, pressure would probably favor the formation of hypersthene. Korjinsky (1937) believes that hypersthene in metamorphic rocks is characteristic of great depth.

By this process the slow differentiation of a magma at depth is necessary to form the potash-rich magma. The tectonic history of the Montana area is favorable to this since in the plains, where most of these rocks are found, the Cretaceous and older rocks are but little deformed, except near the igneous rocks where the deformation is due to the igneous activity.

It is possible that the magmas of the subprovinces of the Montana petrographic province were derived from a basaltic magma by the separation of crystals of plagioclase and hypersthene. The amount of crystals that must be separated is large. Even under the most favorable hypothesis, it is from 55 to 69 per cent of the original magma. If some of the liquid remained between the interstices of the settled crystals, even less liquid would remain. Moreover, some potash would crystallize with the plagioclase and other minerals, and this would complicate the calculations and reduce the amount of liquid available for the potash-rich magmas.

High pressure would favor the formation of garnet and jadeite, but the separation of these minerals would enrich the residual liquid in silica and hence would not yield magmas of the types found in Central Montana but would tend rather toward a granitic magma.

Some alkalic rocks appear to have been formed by the assimilation of limestone by a basaltic magma and the differentiation of this contaminated magma. To form the parent magmas of the petrographic province of Central Montana the early addition of CaO to a basaltic magma would require the separation of a larger amount of crystals, and a larger proportion of the crystals would be diopside and less would be hypersthene and quartz. If 7 per cent of CaO were added to the

basaltic magma to yield a liquid of the composition of the Highwood magma, 76 per cent of crystals of the following composition must be removed:

albite	24	diopside	26
anorthite	37	hypersthene	15

The assimilation of more CaO would convert more of the hypersthene to diopside and olivine. If diopside and olivine are more likely to form early, the assimilation of limestone might favor the formation of magmas of the Highwood type. However there is no reason to believe that this area is more favorable for the assimilation of limestone than many near-by areas of lime-alkalic rocks. The limestone must have been pre-Cambrian, as some of the rocks intrude pre-Cambrian or early Paleozoic rocks. The assimilation must have been highly selective.

The association of an ordinary basaltic magma with the potash-rich magmas and the presence of provinces intermediate in character between the extreme potash-rich magmas and the basaltic magma show a close relation between the types and preclude the interpretation that the potash-rich rocks represent a primary deep-seated magma. It might be possible to derive the various magmas of the subprovinces from a deep-seated layered magma.

Kennedy (1933) and Kennedy and Anderson (1936) propose the hypothesis that lime-alkalic rocks are derived from a magma of the composition of tholeiite (an olivine-poor or olivine-free basalt of the type found in plateau basalts) and that the alkalic rocks are derived from a magma of olivine basalt (of the type found in the oceanic basins). They postulate that the earth's crust is layered and that an olivine basalt layer underlies one of tholeiite. Under orogenic belts the tholeiite is melted, and under nonorogenic belts the olivine basalt is melted. In the Montana province, the apparent gradation between typical basalt-rhyolite subprovinces and typical alkalic subprovinces with no evidence of any sharp distinction between the two types is better explained as due to a common origin for all the subprovinces.

Kennedy's olivine basalt fits the variation curves (projected) for the lime-alkalic provinces much better than it does those for the alkalic provinces, whereas the tholeiite is much farther from curves for lime-alkalic provinces than the olivine basalt. Daly's plateau basalt is between the tholeiite and olivine basalt of Kennedy. As compared with the San Juan curves (projected) the olivine basalt is about 1 per cent low in CaO and 1½ per cent high in FeO, while the tholeiite is 3 per cent low in Al_2O_3, 2 per cent low in MgO, 2 per cent high in SiO_2 and FeO, and 1 per cent high in CaO.

The most probable explanation for the origin of these magmas appears to be by the separation of crystals of pyroxene and plagioclase from a basaltic magma. Some other processes may have played a part.

The BaO and SrO would probably be concentrated with the potash in a process of crystal fractionation such as is here postulated, but probably not to the extent found in the Montana provinces.

DIFFERENTIATION WITHIN THE SUBPROVINCES

General statement.—There is much evidence that the differentiation within the subprovinces was in large part brought about by crystal fractionation. In each subprovince the progressive change in mineral composition from the mafic to the felsic rocks is in large part a change in the proportion of the minerals rather than in their kind, and the change is in the direction that would be expected if the felsic rocks were formed by the removal from the mafic rocks of the minerals which formed early. The increase in the silica content in the felsic rocks, and especially in the change from nepheline syenites to syenites and finally to granites, is an exception to this, and silica must have been added to the magmas.

The progressive change in the mineral composition is well shown by the Highwood subprovince. In this subprovince more than half of the mafic rocks are made up of mafic minerals, chiefly augite and olivine with some biotite. The felsic minerals are chiefly sanidine and nepheline with some leucite and analcime. Olivine was completely precipitated and augite largely precipitated before any large amount of the light constituents were precipitated. Leucite crystallized rather early. Most of the biotite precipitated late. The mafic-poor nepheline syenites have essentially the same minerals as the mafic-rich shonkinites except that they have no olivine, little pyroxene, and considerable biotite. Rarely nepheline syenites contain hornblende.

In the Little Belt subprovince feldspathoids are rare; hornblende is common in shonkinites, syenites, and granites; and plagioclase is present but subordinate to orthoclase in nearly all the rocks. Here, too, the difference between the shonkinites and the syenites is largely a difference in the proportion of mafic minerals. In the quartz syenites and granites quartz is present, and this must require some reaction or the assimilation of silicious material.

The rocks of the Crazy Mountains subprovince are much like those of the Highwood subprovince except that they are richer in Na_2O and poorer in K_2O and that quartz-bearing rocks are present in the Crazy Mountains.

The rocks of the Absaroka subprovince are rich in plagioclase and nearer to the ordinary lime-alkalic rocks.

There is a tendency for the quartz-rich rocks in all the subprovinces to approach a normal granite. The Highwood subprovince has no quartz rocks.

The numerous differentiated laccoliths, stocks, and other bodies in the area were formed by crystal differentiation, and the different rocks of these bodies are like the different rocks of the subprovince to which they belong, indicating that crystal settling is a major process in the differentiation of the magmas of the subprovinces.

The layered laccoliths of the Highwood Mountains were formed by crystal differentiation, as shown by Hurlbut and Griggs (1939). The original magma had the composition of a shonkinite with about 50 per cent of mafic minerals. Chemically and mineralogically these rocks are like the shonkinites of the stocks and the mafic phonolites of the lavas and dikes. The nepheline syenite of the laccoliths was formed by the removal of the mafic minerals from the shonkinitic magma, and it is essentially the same as the phonolites and nepheline syenites of the stocks and the dikes. Nearly the whole range of the rocks of the subprovince is represented in the differentiated laccoliths.

The differentiated stock of Yogo Peak in the Little Belt Mountains contains plagioclase-bearing shonkinite, monzonite, and quartz-bearing syenite. Essentially the same minerals occur in all the rocks but in different amounts. All the rocks contain plagioclase and they are essentially like the other rocks of the Little Belt subprovince.

Quantitative data on the changes in composition within each subprovince.—When we determine the amount and composition of the material that must be removed from the mafic rocks of any subprovince to form the felsic rocks, we find for most subprovinces that the composition is reasonable only after silica is added. The amount of silica that must be added is large for the provinces in which quartz-rich rocks are present, but some may be necessary even in the Highwood subprovince in which the most silicic rock is syenite.

In the following determinations of the material that must be subtracted from one magma to yield another, the method of plotting was used. Unless otherwise stated the values of the oxides were read from the variation diagrams. For convenience in plotting, the position of the rock in the plots was taken as its position on the variation diagram. The calculation can be made directly on the variation diagram. This method ignores the minor constituents and the state of oxidation of the iron.

The Shonkin Sag laccolith offers an unusually good opportunity for a calculation of this kind. The main type of differentiation by crystal settling is established, and a number of good analyses are available. For

this calculation the following analyses were used: (1) the original magma represented by the average of two analyses of the chilled border of the laccolith, (2) the lower shonkinite formed by addition of crystals to the original magma, (3) the transition rock, (4) the syenite, (5) the aegirite

TABLE 7.—*Amounts of material that must have been added to or subtracted from original magma of Shonkin Sag laccolith to yield various rocks of laccolith*

	Composition				
	Added	Subtracted			
	B to A	B to C	B to D	B to E	B to F
SiO_2	45.2	49.5	43.1	43.4	44.3
Al_2O_3	1.5	0	3.1	1.7	3.4
Fe_2O_3	6.0	−.5	.2	1.0	1.9
FeO	10.3	9.4	10.3	9.8	10.0
MgO	25.6	19.5	21.0	22.5	19.2
CaO	12.0	20.1	22.0	20.8	19.5
Na_2O	0	−3.2	−2.1	−2.5	−2.2
K_2O	0	5.4	3.7	3.8	3.3
K_2O less Na_2O deficiency	0	.6	0	0	0
Amount	31	30	25	34	36
Mol. MgO/FeO	6.0	3.9	3.7	4.3	3.8
Amount of theoretical minerals added or subtracted					
K_2SiO_3	.....	.3	.....	.....	.....
an.?	0.3	.....	2.1	1.6	3.4
di	13.6	24.0	20.1	26.9	25.0
hy	3.1	.....	.....	.....	.....
ol	10.3	5.8	5.9	8.7	8.0
mg	2.7	.....	.1	1.5	1.8
SiO_2 deficiency	0	.2	3.3	2.8	2.2

A. Lower shonkinite
B. Average chilled border
C. Transition rock
D. Syenite
E. Aegirite syenite, small intrusive in upper shonkinite
F. Phonolite. Local, late intrusive.

syenite, and (6) the phonolite. For these calculations the TiO_2 was calculated into ilmenite, P_2O_5 into apatite, and BaO and SrO into celsian. These minerals were then subtracted from the original analysis, which was recalculated to 100 per cent. The FeO and Fe_2O_3 were kept separate. The composition and amount of the least amount of material that can be subtracted (or added to yield the lower shonkinite) from the original rock to yield the other rocks are given in Table 7. The amount of theoretical minerals that must have been subtracted or added and the SiO_2 that must have been added to or subtracted from the original

magma to yield the rocks found is also given in Table 7. In making these calculations, a deficiency in Na_2O was deducted in molecular proportions from the K_2O as there is known to have been some paulopost replacement of potash by soda. In the rocks of the Highwood Mountains

TABLE 8.—*Amounts of pyroxene of analyzed type and of other theoretical minerals that must be removed from or added to the original Shonkin Sag magma to yield various analyzed rocks*

	Added	Subtracted			
	B to A	B to C	B to D	B to E	B to F
Augite	17	30	30	32	32
Olivine	14	9	5	8	9
Acmite	½	.....	.....	.....	.....
Fe_2O_3	1	.....	.....	.....	.....
K_2SiO_3	.....	½	½	½	.....
Deficiency:					
Al_2O_3	.....	.....	.....	.....	.....
SiO_2	.....	2	3	1½	2
Amount	32	38	33	40	39

The letters A to F are explained in Table 7.

the nepheline is altered, and there has probably been some replacement of potash by soda.

A more precise calculation, made by using the analyzed pyroxene from Square Butte, is shown in Table 8. In these calculations a deficiency in Na_2O was subtracted from the K_2O and a deficiency in Fe_2O_3 from the FeO after adjusting for molecular replacement. The calculation was made on the assumption that all the CaO goes into pyroxene.

The minerals estimated to have been removed or added in Tables 7 and 8 are very similar if allowance is made for the fact that the pyroxene of Table 8 includes the diopside, anorthite, and magnetite of Table 7.

The minerals that are calculated to have been added to the lower shonkinite (B to A) are chiefly olivine and diopside, and this agrees with the field observations. The same minerals must be taken from the original magma to form the felsic rocks, but some Fe_2O_3 and K_2O must be removed and some SiO_2 added. This SiO_2 probably came from the mineralizers and is in part accounted for by the alteration of the nepheline. The removal of K_2O must be due to the same cause. The removal of biotite would account for some of the loss of K_2O and for a part of the deficiency in silica. The data confirm the conclusion reached by Hurlbut and Griggs that the differentiation of the Shonkin Sag laccolith was by crystal settling, modified somewhat by late mineralizers.

For the Highwood subprovince the composition and least amounts of material that must be successively removed to make the succeeding magmas move along the variation diagrams are shown in Table 9. The amounts of calculated minerals that must be removed are also shown.

TABLE 9.—*Composition and least amounts of material that must be removed from the magmas of the Highwood subprovince to move the chemical composition along the variation diagram of the subprovince, and calculated minerals that must be removed*

Move from position to	−15 to −10	−10 to 0	0 to 10	10 to 15	15 to 20
SiO_2	45.2	42.7	45.8	36.0	35.6
Al_2O_3	0	3.7	6.0	0	27.6
Fe_2O_3	0	6.7	2.1	20.2	4.4
FeO	13.9	5.0	9.1	1.2	10.0
MgO	26.5	25.5	15.0	16.5	4.5
CaO	12.0	16.1	19.4	24.2	13.7
Na_2O	0.7	0	0	0.5	3.0
K_2O	2.5	0	0	0.2	0
Amount	16½	25	23½	8	12
Calculated minerals to be subtracted					
Nepheline					1.6
K_2SiO_3	.4				
Acmite				1.1	
Anorthite		2.5	3.9		7.4
Diopside	8.0	13.2	15.0	6.9	.6
Olivine	8.2	7.5	3.6	.1	2.1
Magnetite		2.4	.7	1.1	.8
Hematite				.4	
SiO_2 deficiency	0	1.0	.5	1.5	.7
MgO / FeO mol	3.5	24.0	3.3	0	1.0
Percentage of original magma remaining	84	63	48	45	39

This calculation was made in the same way as that for the Shonkin Sag laccolith in Table 7. If the minor constituents are not taken into account and the iron is all calculated as FeO, nearly the same results are obtained. The last column is not satisfactory, as there are few analyses in this range and they have a variable content of Al_2O_3.

The minerals that it is necessary to remove from the mafic rocks to form the felsic rocks and stay on the variation curves are chiefly olivine and diopside, minerals that we would expect to crystallize from the magma.

The pyroxene would be essentially like that of Square Butte laccolith. If the calcium to be deducted is assumed to go into such a pyroxene, the minerals to be deducted during each stage are shown in Table 10.

TABLE 10.—*Pyroxene like that analyzed from Square Butte, and other calculated minerals to be deducted to change the magma of the Highwood subprovince along the variation curve*

	−15 to −10	−10 to 0	0 to 10	10 to 15
Pyroxene	10	18	20	9
Olivine	8	7	4	½
K_2SiO_3	1			
Acmite				2½
Al_2O_3			½	
Fe_2O_3		1		
SiO_2 added	0	1¼	1½	1½

TABLE 11.—*Least amounts of material that can be removed from the mafic rocks of the Crazy Mountains subprovince to yield the felsic rocks as calculated from the variation diagram*

Chemical composition				Percentage of theoretical minerals removed			
Positions	−5 to 5	5 to 15	15 to 23	Positions	−5 to 5	5 to 15	15 to 23
SiO_2	0	26.2	38.0	or			9.8
Al_2O_3	1.2	0	26.4	lc	3.2	2.5	
FeO	15.0	28.8	12.6	ne	8.4	.2	
MgO	21.8	18.1	4.2	an			15.3
CaO	43.6	21.2	12.2	di	22.7	9.6	1.4
Na_2O	13.5	.3	0	$CaSiO_3$	.4		
K_2O	5.2	4.9	5.9	hy		6.0	8.6
Mol. MgO/FeO	2.6	1.1	.6	ol		4.4	
Amount	13½	11	28				
Percentage of original magma (at −5) remaining	86	77	56	Add Al_2O_3	3.6	.6	
				SiO_2	17.7	6.7	7.3

This is very much like the preceding calculation with the anorthite and magnetite added to the pyroxene.

The minerals shown in Table 9 and better in Table 10 are quantitatively essentially the minerals that, from the field and laboratory study, we could expect to settle from the mafic magmas. The small amount of silica that must be added is believed to be in part due to the alteration of the nepheline and in part to a process related to gas transfer, in part to the removal of biotite.

TABLE 12.—*Least amounts of material that can be removed from the mafic rocks of the Little Belt subprovince to yield the felsic rocks as calculated from the variation diagram*

Chemical composition					Percentage of theoretical minerals removed				
Positions	−10 to 0	0 to 10	10 to 20	20 to 29	Positions	−10 to 0	0 to 10	10 to 20	20 to 29
SiO_2	32.0	0	0	0	or	.....	2.4	1.2	4.9
Al_2O_3	8.1	2.2	19.2	35.2	ab	.....	1.0	.4	4.2
FeO	16.4	27.6	25.0	21.4	ne	1.2	.....	.....	.....
MgO	20.8	34.0	26.6	11.6	an	4.0	.....	4.4	6.6
CaO	21.4	31.2	26.6	21.0	di	15.7	12.6	7.2	4.8
Na_2O	1.2	1.2	.5	4.4	hy	.....	7.5	7.7	5.9
K_2O	0	4.0	2.0	6.8	ol	8.0	13.0	.....	.....
Add leucite	.....	.....	4.0	.....	Add				
Mol. MgO/FeO	2.3	2.2	1.9	1.0	SiO_2	6.7	13.0	10.9	14.6
Amount	22½	10	10	12	lc	.....	.....	3	
Percentage of original magma (at −10) remaining	77	70	63	55					

TABLE 13.—*Least amounts of material that can be removed from the mafic rocks of the Absaroka subprovince to yield the felsic rocks as calculated from the variation diagram*

Chemical composition					Percentage of theoretical minerals removed				
Positions	−10 to −2	−2 to −1	−1 to 10	10 to 20	Positions	−10 to −2	−2 to −1	−1 to 10	10 to 20
SiO_2	36.0	.....	47.0	16.5	or	.....	3.3	3.5	1.3
Al_2O_3	2.0	.....	9.9	35.2	lc	1.2	.....	.....	.....
FeO	9.9	.....	15.2	19.4	ab	.....	4.0	.....	.....
MgO	38.0	.....	10.5	13.0	cor	.....	.....	.....	1.1
CaO	13.0	.....	16.4	15.0	an	.2	.....	7.0	12.3
Na_2O	0	.....	0	0	di	8.3	.....	16.2	.....
K_2O	1.5	.....	1.8	1.4	hy	.....	3.1	9.7	11.1
Mol. MgO/FeO	7.0	.....	1.3	1.6	ol	10.6	7.5	.....	.....
Amount	16⅔	18	33	16	Add SiO_2	3.6	0	3.6	9.3
Percentage of original magma (at −10) remaining	83	68	45	38					

All the evidence from the field and laboratory study of the Highwood subprovince shows that crystal fractionation was the chief cause of the differentiation, modified slightly by the concentration of silica in the felsic rocks by gas transfer.

Calculations similar to those made from the variation diagrams for the Highwood subprovince (Table 9) have been made for the other subprovinces and are given in Tables 11 to 16.

TABLE 14.—*Least amounts of material that can be removed from the mafic rocks of the Crandall subprovince to yield the felsic rocks as calculated from the variation diagram*

Chemical composition				Percentage of theoretical minerals removed			
Positions	−7 to 0	0 to 5	5 to 16	Positions	−7 to 0	0 to 5	5 to 16
SiO_2	50.2	33.0	31.0	or	3.1	1.6	.....
Al_2O_3	10.3	11.9	27.4	ab	3.1	.....	5.3
FeO	8.2	15.6	19.4	an	7.2	3.1	12.3
MgO	16.0	17.4	9.4	cor	.....	.....	.7
CaO	13.7	20.0	10.8	di	11.2	7.2	.....
Na_2O	1.2	0	2.7	hy	3.1	5.2	13.6
K_2O	0	2.3	0	ol	6.2	.....	.....
Mol. MgO/FeO	3.5	2.0	.9	Add SiO_2	.....	5.0	8.7
Amount	31	12	23				
Percentage of original magma (at −7) remaining	69	61	47				

TABLE 15.—*Least amounts of material that can be removed from the mafic rocks of the Yellowstone subprovince to yield the felsic rocks as calculated from the variation diagram*

Chemical composition						Percentage of theoretical minerals removed					
Positions	−10 to 0	0 to 10	10 to 20	20 to 25	25 to 29	Positions	−10 to 0	0 to 10	10 to 20	20 to 25	25 to 29
SiO_2	46.2	42.2	0	30.0	71.7	Q	.....	.....	.....	.....	17.2
Al_2O_3	14.9	18.8	20.0	34.0	17.3	cor	.....	.....	.....	.....	3.1
FeO	13.0	13.7	26.0	7.0	3.0	or	.....	.....	.6	.....	.....
MgO	10.3	10.0	22.5	6.0	.6	ab	6.0	7.0	1.7	4.9	18.3
CaO	13.4	12.1	27.9	15.5	1.4	an	14.7	13.3	4.8	3.9	3.2
Na_2O	1.6	2.5	1.8	8.2	4.7	di	12.0	5.3	8.6	1.2	.....
K_2O	0	0	.8	0	0	hy	.....	14.0	7.2	1.4	3.2
Mol. MgO/FeO	1.4	1.3	1.6	1.6	.4	ol	11.7	.....	.....	.....	.....
Amount	44	33⅓	11	7	46	Add SiO_2	.6	6.7	11.9	4.3	.....
Percentage of original magma (at −10) remaining	56	37	33	31	16						

Discussion of the data.—The data of Tables 9 to 16 are plotted on Figure 20. In the left-hand column are cumulative curves which show the percentage of theoretical minerals whose removal from the magma on the extreme left yields the magma at any point. The right-hand column shows the rate at which the theoretical minerals are removed at any stage. It shows the percentage of minerals, considering the magma at that point as 100 per cent, which must be removed to move the magma 10 points along the variation curve.

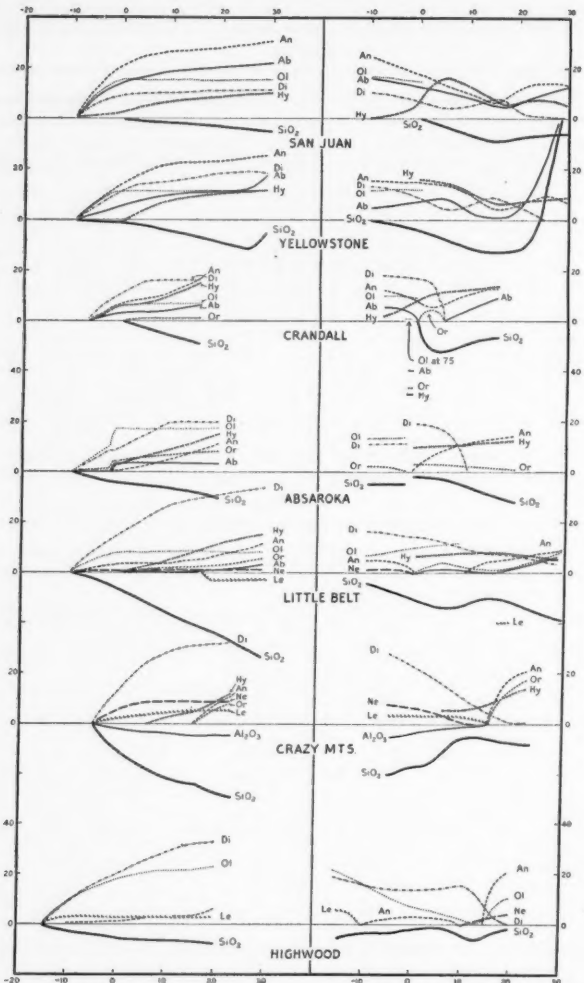

FIGURE 20.—*Diagram to show the course of removal of theoretical minerals from the magmas of the various petrographic subprovinces*

Left-hand column shows percentage of theoretical minerals whose removal at the extreme left yields the magma at any point. Right-hand column shows rate at which minerals are removed at any stage in terms of percentages of magma at that point to move the new magma ten points along the variation curves.

In Table 17 are listed the chief features of the diagrams. This table and the diagrams show that from 41 to 86 per cent of material must be removed from the mafic magmas to yield the felsic magmas. The amounts are higher in lime-alkalic provinces than in the alkalic provinces.

TABLE 16.—*Least amounts of material that can be removed from the mafic rocks of the San Juan province to yield the felsic rocks as calculated from the variation diagram*

Chemical composition					Percentage of theoretical minerals removed				
Positions	−10 to 0	0 to 10	10 to 20	20 to 30	Positions	−10 to 0	0 to 10	10 to 20	20 to 30
SiO_2	48.8	48.6	34.3	38.9	ab	14.5	10.5	6.8	9.2
Al_2O_3	16.7	17.9	19.5	29.3	an	21.0	13.9	7.0	13.5
FeO	12.1	12.2	18.6	11.0	di	9.2	5.4	6.7	.2
MgO	9.1	7.3	8.2	3.8	hy	2.2	13.5	7.5	6.8
CaO	10.3	10.3	15.1	12.0	ol	15.5	.....	.....	.....
Na_2O	2.7	3.1	4.0	4.6	Add SiO_2	0	3.6	8.0	6.6
K_2O	0	0	0	0					
Mol. MgO/FeO	1.4	1.1	.7	.6					
Amount	63	40	20	26					
Percentage of original magma (at −10) remaining	37	23	19	14					

In the San Juan and other lime-alkalic provinces the chief phenocrysts found in the rocks from positions −10 to zero are plagioclase, olivine, and augite; from zero to 10 plagioclase, hypersthene, and augite; from 10 to 20 plagioclase, augite, hornblende, and biotite; and to the right of 20 chiefly plagioclase, hornblende, and biotite. At the extreme right, quartz, orthoclase, and biotite are the chief phenocrysts. In the Highwood province olivine crystallized in rocks from −15 to zero. Leucite crystallized near zero. Augite crystallized throughout, and it contained acmite in the highly felsic rocks. Biotite crystallized in small amount throughout but for the most part it crystallized very late. Some early sanidine phenocrysts are present.

The minerals that crystallized from the magmas differ considerably in chemical composition from the calculated minerals, as is shown in Table 18.

The degree to which the curves must be modified to fit the actual minerals of the rocks can be calculated with a fair degree of accuracy for the San Juan province and the Highwood subprovince, both of which are well known to the author and for which some analyses and other

data are available for the minerals. A comparison of the calculated and actual minerals for the San Juan province is shown in Table 19. That for the Highwood province has already been given (Tables 9 and 10).

TABLE 17.—*Chief features of the diagrams for the various petrographic provinces*
Showing the theoretical minerals removed from the mafic magmas to yield the felsic magmas.

	Amount removed			Minerals removed			Silica deficiency		
	−10 to 0	−10 to 20	−10 to 30	Chief	Much	Some	at 20	at 30	
San Juan	63	81	86	calcic plag.	ol. di hy		3½	5	Femic decrease toward right.
Yellowstone	44	67	85	calcic plag.	di ol hy		10	4	Q and ab lost in large amount at extreme right.
	−7 to 0								
Crandall (projected)	31	58		calcic plag. di	hy ol	or	10		Di chief mineral removed at left part.
Absaroka	30	62		di ol	hy Very calcic plag.	or	10		Break in curve.
Little Belt	23	37	43	di	hy Very calcic plag. ol	or ab ne	24	35	
	−5 to 5	−5 to 23					at 23		
Crazy Mountains (projected)	14	44		di	ne lc hy	an or	30		Al_2O_3 deficiency in mafic range.
	−15 to 0	−15 to 20					at 25		
Highwood	37	61		di ol		an lc ne	7		Di is dominant nearly to right end.

There is a silica deficiency in every subprovince, and the maximum amount is from 5 to 10 per cent in the lime-alkalic provinces, from 30 to 35 per cent in the alkalic provinces that contain granite, and only 7 per cent for the Highwood province, in which no quartz-bearing rocks have been found.

In some provinces small amounts of Al_2O_3 or leucite appear to have been added. In the Yellowstone province SiO_2 appears to have been added over most of the range, but it was subtracted in large amount at

TABLE 18.—*Analyzed pyroxenes, amphiboles, and biotite of the San Juan and Highwood areas*

Calculated to the theoretical minerals used in Figure 30 and Table 25.

	San Juan					Highwood
	Hypersthene from andesite	Pyroxene from andesite	Pyroxene from andesite-latite	Amphibole from quartz latite	Biotite from quartz latite	Pyroxene from Shonkinite
or	.....	.....	.....	5.8	54.2	2.2
ab	.....	5.9	2.6	12.0	7.1	8.4
an	4.9	5.3	2.8	12.4	9.3	6.4
di	3.8	68.5	83.6	37.3	3.6	89.2
hy	93.5	25.2	14.0	47.3	65.0	5.1
Mol. MgO/FeO	1.8	2.1	2.5	1.5	2.0	3.0
SiO_2 deficiency	2.1	5.0	2.8	14.8	38.9	11.4

TABLE 19.—*Comparison of calculated and actual minerals that would separate from the San Juan magmas*

	0 to 10		10 to 20		20 to 30	
	Calculated	Actual	Calculated	Actual	Calculated	Actual
Clinopyroxene	5.4	7.0	6.7	5.1	.2	.....
Orthopyroxene	13.5	11.0	7.5	2.3	6.8	.....
Amphibole	.....	1.1	.....	7.9	.....	1.0
Biotite	.....	1.4	.....	2.0	.....	13.0
Anorthite	13.9	13.0	7.0	6.0	13.5	12.5
Albite	10.5	9.6	6.8	5.7	9.2	8.2
SiO_2 deficiency	3.6	2.6	8.0	6.2	6.6	2.4

the extreme felsic end, where quartz and sodic feldspar appear to have separated. This process would lead to a potash granite. In the Highwood subprovince the silica deficiency can be accounted for by subtracting the actual minerals of the rocks rather than the theoretical minerals. In the San Juan province and in the Yellowstone, Crandall, and Absaroka subprovinces the silica deficiency is too large to be accounted for in this way, and siliceous material was probably assimilated.

In the Little Belt and Crazy Mountains subprovinces the silica defi-

ciency is much too large to be explained as due to a silica deficiency in the actual minerals that crystallized. Assimilation of siliceous material in large amounts is necessary. This assimilated material was probably of granitic composition.

Cause of the variation in composition within the subprovinces.—The regularity of the curves of the variation diagrams for which abundant data are available shows that some rather simple process must control the variation of the rocks. The failure of a moderate number of analyses to fit the variation curves is in part due to poor analyses, poor sampling, or alteration, but it is probably in part due to some interference with the normal course of the process of variation.

In the Highwood subprovince the rocks of the lavas, dikes, and stocks are like those of the laccoliths which have been shown to have formed by crystal settling, and the conclusion seems to be justified that the differentiation in this subprovince was chiefly by crystal sorting. In the San Juan province the compositions of the groundmasses of the rocks, many of which contain from 25 to 50 per cent of the phenocrysts, fall on the same variation diagram as do the rocks. The groundmasses are displaced toward the felsic end by an amount proportional to the content of phenocrysts. Moreover, it has been shown that the variation within all the subprovinces of the Central Montana petrographic province can be accounted for by starting with a rock near the mafic end of the diagrams and removing by stages the minerals that would crystallize from the magma at that stage and adding silica.

A serious objection to such a process of magmatic differentiation is the small amount of felsic rocks that are formed. In the San Juan province, according to the calculations, no K_2O was removed at any stage. The settled crystals and the interstitial liquid would certainly contain some. In the typical basalts of that province all the K_2O goes into the plagioclase and other minerals. This K_2O in the settled crystals and interstitial liquid would reduce the amount of the felsic differentiate. In the San Juan lavas, after sufficient crystals are removed so as to change the magma from basalt (at —10) to pyroxene andesite (at 0), only 37 per cent of the latter is formed. In changing the magma from basalt to hornblende andesite (at 10), only 23 per cent of andesite is formed, and to move to the extreme rhyolite (at 30) only 14 per cent of rhyolite is formed. These are the maximum amounts. In the Yellowstone subprovinces the amounts are somewhat larger as shown in Table 17. In the alkalic provinces, where silica is added, the amounts are larger. The assimilation of granitic material to furnish the necessary SiO_2 would add to the amount of felsic differentiate.

The variation diagram for a mixture in various proportions of a mafic rock and granite would yield a variation diagram with straight lines. The actual variation diagrams of all the provinces are made up of curves but with such gentle curvature in most parts that a considerable amount of granite could be added to a magma represented by a point on the diagram and yet yield a mixture that would fall on the diagram within the limits of error of the data.

Evidence that rocks do assimilate foreign rock is shown by the fact that some of the phenocrysts of the San Juan lavas are foreign to the rock (Larsen, Irving, Gonyer, and Larsen 3d, 1938, p. 429). In the San Luis Rey quadrangle of California, Hurlbut (1935) found in a tonalite numerous inclusions of gabbro which show all stages of reaction to complete mixing with the tonalite. Partly resorbed siliceous inclusions are much less common than mafic ones, probably because siliceous inclusions are more readily reacted on and dissolved by the magma.

Kennedy (1933) states that in the olivine basalt, which is the parent magma for alkalic rocks, crystallization begins well in the olivine field and yields lime-rich pyroxene. The magma is impoverished in CaO. In tholeiite, which is the parent magma of the lime-alkalic rocks, crystallization begins in or near the pyroxene field and yields lime-poor pigeonite, and the magma increases in CaO.

Figure 21 shows the proportions of FeO, MgO, and CaO and the change in their proportions along the variation curves for the petrographic subprovinces discussed in this paper. For various positions on the variation diagrams the oxides were read and their sum calculated to 100 per cent. This yields fairly accurate plots for the mafic rocks but only rough approximations for the felsic rocks, in which the total of the three oxides is only a few per cent. This figure shows a more rapid increase in the relative amount of CaO over most of the range in the lime-alkalic subprovinces than in the alkalic subprovinces.

In the lime-alkalic provinces of Yellowstone and San Juan MgO decreases rapidly from basalts to quartz latites, CaO increases rapidly, and FeO increases slightly. At the quartz latites there is an abrupt break in the curves, and from quartz latites to mafic-poor rhyolites CaO decreases rapidly, and FeO increases.

In the alkalic subprovinces (Highwood, Little Belt, and Absaroka) MgO is lower and FeO higher in the mafic rocks than in the lime-alkalic provinces. There is a small increase in CaO and a large increase in FeO at the mafic end and an increase in FeO at the felsic end. In the Absaroka province the large increase in MgO near the mafic end is caused by the break in the curves for MgO and CaO in the variation diagram of that province.

Conclusion.—The preceding data and discussion lead to the highly probable conclusion that the variation in the rocks of the petrographic provinces discussed was caused in large part by crystal differentiation but to a greater or less extent by the assimilation by reaction on granitic or

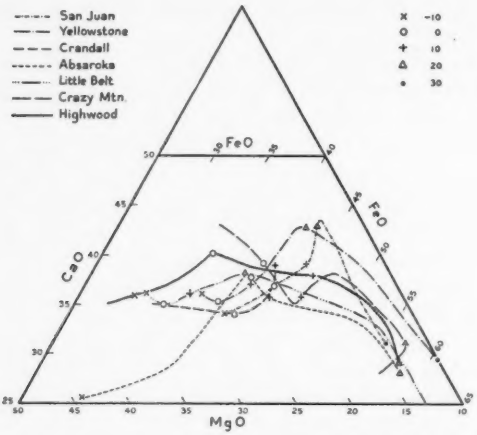

FIGURE 21.—*Triangular diagram to show relative proportions of FeO, MgO, and CaO for the subprovinces* Central Montana and the San Juan Mountains of Colorado.

other siliceous material. Gas transfer and other processes may have played a minor part. In the Highwood subprovince little or no assimilation is required, in the San Juan province a small amount of assimilation is probable, in the other subprovinces assimilation was more important and the amount of material assimilated in the Little Belt and Crazy Mountains subprovinces must have been nearly as large as the amount of crystals removed. The assimilation took place with the precipitation of crystals and in this way no superheat was required, as the latent heat necessary for assimilation was furnished by precipitation of the crystals.

The assimilation of granitic material by the magmas would increase the amount of felsic differentiate and thus yield an adequate amount.

GENERAL SUMMARY AND CONCLUSIONS

The petrographic province of Central Montana is rather arbitrarily defined to include the Tertiary and Quaternary igneous rocks along or near the eastern base of the Rocky Mountains from the Yellowstone National Park on the south to the Canadian boundary on the north. These igneous

rocks are scattered over an area nearly 400 miles long and between 100 and 200 miles wide. There are about 12 large mountain masses formed chiefly by the igneous rocks and many smaller igneous bodies. Six of the larger areas have extrusive rocks. Large laccoliths are characteristic of the alkalic parts of the province, and stocks, sills, and dikes are numerous. Each of the areas represents rocks originating from one or several centers of eruption.

The rocks of the province range from ordinary rhyolites, andesites, and basalts to potash and soda-rich rocks—shonkinites, nepheline seynites, and syenites—that are almost devoid of plagioclase. The rocks of each mountain group fall into one or more subprovinces, and the rocks of each subprovince have peculiar mineral and chemical features, although they commonly range from highly mafic to highly felsic rocks. Chemically the rocks of a subprovince fall on or near a variation diagram. In some of the mountain ranges as many as four subprovinces have been recognized, and the rocks of the several subprovinces are very different. Where data are available, subprovinces represent eruptions within a time interval, and the several subprovinces are separated by intervals with few or no eruptions, as shown by extensive erosion. The order of eruptions in the alkalic subprovinces seems to be from mafic to felsic types.

The characteristic features of the chief subprovinces are listed in Table 20. The rocks are compared with the rocks of the San Juan Mountains, which is taken as a typical lime-alkalic province.

In each of the subprovinces a rock near the mafic end is believed to represent the primary magma of the subprovince. This rock ranges in the different subprovinces from an ordinary basalt to orthoclase basalt, to plagioclase shonkinite, and to shonkinite rich in potash and lacking plagioclase. This gradational character of the subprovinces and their close association in time and space lead to the conclusion that they have a common origin and are parts of a larger petrographic province.

This requires two periods of magmatic differentiation, the first a deep-seated differentiation to yield the primary magmas of the subprovinces and then a shallower differentiation of these primary magmas to yield the various rocks of each subprovince.

The primary magmas of the subprovinces are believed to have been derived from a basaltic magma by the removal of crystals of calcic plagioclase and hypersthene in depth. To form the alkalic rocks a long period of quiet differentiation in depth would be required. The relatively flat-lying Cretaceous rocks of the plains area indicate a long period without much deformation, and this would favor slow differentiation.

The variation in the rocks of each subprovince is explained as due to crystal settling combined with the assimilation of siliceous (granitic) ma-

TABLE 20.—*Characteristic features of the subprovinces of the petrographic province of Central Montana*

Subprovince	Chief rocks	Characteristic minerals	Chemistry	Rocks included	Remarks
Yellowstone Area					
1. Yellowstone	Basalts, andesites, rhyolites		SiO_2 high K_2O low	All rocks except lower basic flows and possibly lower basic breccia and associated intrusives	Basal part is Yellowstone type overlain by Crandall, then by Absaroka, and this by main Yellowstone rocks
2. Crandall	Basalts, andesites, rhyolites		Near average lime-alkalic rocks	Crandall volcano and possibly lower basic breccia	
3. Absaroka	Orthoclase basalts, etc., and intrusives	Orthoclase in basalts and gabbros. Rare leucite	K_2O a little high SiO_2 a little low	Lower basic sheets and intrusives	Erosion separates subprovinces
Crazy Mountains					
1. Stocks, etc., of southern area	Diorite, etc.		Lime-alkalic	Stocks and dikes	
2. Crazy Mountains northern area	Shonkinite, nepheline syenite, to granite	Pyroxene, orthoclase, nepheline, quartz	Na_2O very high K_2O high SiO_2 and FeO low CaO and MgO high	Laccoliths, sills, stocks, and dikes	Younger than diorite
Little Belt and Castle Mountains					
1. Diorite	Diorite	Hypersthene	Lime-alkalic	Stocks in both mountains	
2. Little Belt	Plagioclase shonkinite to granite	Orthoclase > plagioclase. Some hornblende and quartz	K_2O high Na_2O rather high Ca and Mg high Fe and Al low	Some lava, laccoliths, stocks	
Highwood Mountains					
1. Quartz latite	Quartz latite		K_2O rather high. Near Little Belt or Absaroka	Large volcanic mountain	Quartz latite is older than Highwood subprovince and separated by extensive erosion
2. Rocks of Highwood Peak	Quartz syenite, monzonite	Some plagioclase and quartz	Near Little Belt	Single stock	
3. Highwood	Shonkinite, nepheline syenite	Leucite. Trace of plagioclase, no quartz, little hornblende	K_2O very high CaO and MgO high SiO_2 very low	Later volcanic rocks and most of intrusives Stocks, sills, dikes, and laccoliths	

4. Alnoite	Alnoite, monticellite peridotite		CaO very high SiO_2 very low	Single small intrusive	
Bearpaw Mountains					Order of age shown by 1 to 5
1. Lime-alkali types	Quartz monzonite to quartz syenite. Basalt and latite	Quartz, orthoclase, plagioclase	SiO_2 high	Sills, dikes, small stocks, and widespread volcanic flows and breccia	Epoch followed by erosion. Angular inclusions of dunite, peridotite, pyroxenite and gabbro
2. Alkalic	Peridotite, mica pyroxenite, nepheline or pseudoleucite shonkinites, and syenites	Fresh nepheline or fresh pseudoleucite. Few types with labradorite	SiO_2 low K_2O, CaO, MgO high Na_2O not so high as would be expected	Part of composite stocks, sills, dikes, and laccoliths	
3. Lime-alkalic types	Mafic monzonites to plagioclase syenites	Hornblende, olivine, some aegirine-augite, fine-grained andesine	SiO_2, CaO, MgO high	Several widespread simple stocks, and also part of composite stocks	
4. Alkalic, volatile rich	Nepheline and soda syenites	Nepheline, sanidine, aegirite	SiO_2 low K_2O high TiO_2, BaO high	Part of composite stocks	Unusual pegmatites and veins
5. Alkalic	Pseudoleucite tinguaite porphyry, mafic to salic leucite volcanic rocks	Fresh pseudoleucite in dikes, fresh leucite in flows	SiO_2 low K_2O high	Dikes in and near composite stocks. Flows widespread	
Big Belt Mountains near Absaroka subprovince					Little known extrusives and intrusives
Sweet Grass Hills	Diorite to quartz syenite	Plagioclase, quartz		Stocks, some potash-rich intrusives	
Little Rocky Mountains	Granite porphyry with plagioclase syenite			Stock	
Judith Mountains	Quartz porphyry, monzonite, syenite			Laccolith	
South Mocassin Mountains	Rhyolite porphyry			Laccolith	
North Mocassin Mountains	Quartz syenite			Laccolith	

terial. The amount of assimilated material is small for the lime-alkalic subprovinces, absent in the Highwood subprovince in which there are no quartz rocks, and large in the Crazy Mountains and Little Belt subprovinces, in which syenites are followed by granites.

TABLE 21.—*Chief minerals removed from the mafic magmas of the various subprovinces to move the resulting magma toward the felsic rocks*

	Basalt to andesite	Andesite to quartz latite	Quartz latite to rhyolite	Rhyolite to potash rhyolite
Lime-alkalic San Juan Yellowstone Crandall	olivine plagioclase augite	plagioclase augite hypersthene amphibole	plagioclase biotite hornblende	sanidine or sodic plagioclase biotite quartz
	Orthoclase gabbro or plagioclase shonkinite to monzonite	Monzonite to syenite	Syenite to granite	
Plagioclase-orthoclase alkalic types Little Belt Absaroka	augite olivine some plagioclase	augite olivine plagioclase biotite	plagioclase augite hornblende biotite	
	Shonkinite to mafic nepheline syenite	Mafic nepheline syenite to syenite	Syenite to granite	
Soda shonkinite Crazy Mountains	augite nepheline leucite olivine (?)	augite nepheline (?) leucite	diopside nepheline (?) biotite plagioclase	
		Shonkinite to syenite		
Shonkinite Highwood		diopside olivine biotite (?)		

The chief minerals that are believed to have been removed from the primary magmas of the various subprovinces as the magmas changed from mafic to felsic are listed in order of abundance in Table 21. The material listed as removed is modified somewhat by the material added. This may account for some of the nepheline and leucite listed as removed from the Crazy Mountains subprovince.

WORKS TO WHICH REFERENCE IS MADE

Blixit, J. E. (1933) *Geology and gold deposits of the North Moccasin Mountains, Fergus County, Montana,* Mont. Bur. Mines Geol., Mem. 8, p. 6-10.

Daly, R. A. (1933) *Igneous rocks and the depths of the earth.* McGraw-Hill Book Company, New York.

Emmons, W. H. (1908a) *Geology of the Haystack Stock, Cowles, Park County, Montana,* Jour. Geol., vol. 16, p. 193-229.

——— (1908b) *Gold deposits of the Little Rocky Mountains, Montana,* U. S. Geol. Survey, Bull. 340, p. 96-116.

Fenner, C. N. (1926) *The Katmai magmatic province,* Jour. Geol., vol. 34, p. 673-772.

Hague, Arnold (1899) *Absaroka, Montana,* U. S. Geol. Survey, Geol. Atlas, Folio no. 52.

———, and Iddings, J. P. (1896) *Yellowstone National Park, Wyoming,* U. S. Geol. Survey, Geol. Atlas, Folio no. 30.

———, ———, and Weed, W. H. (1899) *Geology of the Yellowstone National Park, Part 2, Descriptive geology, petrography and paleontology,* U. S. Geol. Survey, Mem. 132, p. 1-439.

Hurlbut, C. S. (1935) *Dark inclusions in a tonalite of Southern California,* Am. Mineral, vol. 20, p. 609-630.

———, and Griggs, David (1939) *Igneous rocks of the Highwood Mountains, Montana; Part 1, The laccoliths,* Geol. Soc. Am., Bull., vol. 50, p. 1043-1112.

Iddings, J. P., and Weed, W. H. (1894) *Livingstone, Montana,* U. S. Geol. Survey, Geol. Atlas, Folio no. 1.

Kemp, J. F., and Billingsley, Paul (1921) *Sweet Grass Hills, Montana,* Geol. Soc. Am., Bull., vol. 32, p. 437-478.

Kennedy, W. Q. (1933) *Trends of differentiation in basaltic magmas,* Am. Jour. Sci., 5th ser., vol. 25, p. 239-256.

———, and Anderson, E. M. (1936) *Crustal layers and the origin of magmas,* Bull. volcanologique, ser. 2, tome 3, p. 23-82.

Korjinsky, D. W. (1937) *Dependence of mineral stability on depth,* Soc. Min. Russe, Mem. 3, Lib. 2, p. 385.

Larsen, E. S. (1938) *Some new variation diagrams for groups of igneous rocks,* Jour. Geol., vol. 46, p. 506-520.

———, Irving, John, Gonyer, F. A., and Larsen, E. S. 3rd (1936) *Petrologic results of a study of the minerals from the Tertiary volcanic rocks of the San Juan region, Colorado,* Am. Mineral, vol. 21, p. 677-701;
Same (1937) Am. Mineral; vol. 22, p. 889, 905.
Same (1938) Am. Mineral; vol. 23, p. 227-257 and 417-429.

Lindgren, Waldemar (1886) *Eruptive rocks of Montana,* U. S. 10th Census, vol. 15, p. 719-737.

——— (1891) *Eruptive rocks of Montana,* Calif. Acad. Sci., vol. 3, p. 39-57.

——— (1893) *A sodalite syenite and other rocks from Montana,* Am. Jour. Sci., 3d ser., vol. 45, p. 286-297.

Palmer, H. S. (1925) *Structure of the South Moccasin laccolith, Fergus County, Montana,* Am. Jour. Sci., 4th ser., vol. 10, p. 119-133.

Parsons, W. H. (1939) *Volcanic centers of the Sunlight area, Park County, Wyoming,* Jour. Geol., vol. 47, p. 1-26.

Pirsson, L. V. (1905a) *Petrography and geology of the igneous rocks of the Highwood Mountains, Montana,* U. S. Geol. Survey, Bull. 237.

——— (1905b) *The petrographic province of Central Montana,* Am. Jour. Sci., 3d ser., vol. 20, p. 35-49.

Reeves, Frank (1927) *Geology of the Cat Creek and Devil's Basin oil fields and adjacent areas in Montana,* U. S. Geol. Survey, Bull. 786, p. 39-95.

Reeves, Frank, (1931) *Geology of the Big Snowy Mountains, Montana,* U. S. Geol. Survey, Prof. Paper 165, p. 135-149.

Ross, C. S. (1926) *Nepheline-hauynite alnoite from Winnett, Montana,* Am. Jour. Sci., 5th ser., vol. 11, p. 218-227.

Rouse, J. T. (1933) *The structure, inclusions and alteration of the Deer Creek intrusive, Wyoming,* Am. Jour. Sci., 5th ser., vol. 26, p. 144.

——— (1935) *The volcanic rocks of the Valley area, Park County, Wyoming,* Am. Geophys. Union, Tr., 16th Ann. Meeting, pt. 1, p. 274-284.

——— (1937) *Genesis and structural relationships of the Absaroka volcanic rocks, Wyoming,* Geol. Soc. Am., Bull., vol. 48, p. 1257-1295.

Rouse, J. T., Hess, H. H., Foote, F., Vhay, J. S., and Wilson, K. P. (1937) *Petrology, structure and relation to tectonics of porphyry intrusions in the Beartooth Mountains, Montana,* Jour. Geol., vol. 45, p. 717-740.

Weed, W. H. (1899a) *Fort Benton, Montana,* U. S. Geol. Survey, Geol. Atlas, Folio no. 55.

——— (1899b) *Little Belt Mountains, Montana,* U. S. Geol. Survey, Geol. Atlas, Folio no. 56.

———, and Pirsson, L. V. (1895a) *On the rocks of the Sweet Grass Hills, Montana,* Am. Jour. Sci., 3d ser., vol. 50, p. 309-313.

———, ——— (1895b) *Igneous rocks of Yogo Peak, Montana,* Am. Jour. Sci., 3d ser., vol. 50, p. 467-479.

———, ——— (1895c) *Highwood Mountains of Montana,* Geol. Soc. Am. Bull., vol. 6, p. 389-422.

———, ——— (1896a) *The Bearpaw Mountains, Montana,* Am. Jour. Sci., 4th ser., vol. 1, p. 283-301, 351-362; vol. 2, p. 136-148, 188-199.

———, ——— (1896b) *The geology of the Little Rocky Mountains, Montana,* Jour. Geol., vol. 4, p. 399-428.

———, ——— (1896c) *Geology of the Castle Mountains mining district, Montana,* U. S. Geol. Survey, Bull. 139.

———, ——— (1897) *Geology and mineral resources of the Judith Mountains, Montana,* U. S. Geol. Survey, Ann. Rept. 18, pt. 3, p. 437-616.

———, ——— (1899) *Geology of the Little Belt Mountains, Montana,* U. S. Geol. Survey, Ann. Rept. 20, pt. 3, p. 257-581.

Wolff, J. E. (1938) *Igneous rocks of the Crazy Mountains, Montana,* Geol. Soc. Am., Bull., vol. 49, p. 1569-1628.

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CORRECTIONS AND EMENDATIONS

Contributors to Volume 51 have been invited to send corrections and emendations to be made in their paper, and the volume has been scanned with some care. Corrections and insertions are as follows:

Page 1361, line 19, *should read* a 50-ton treatment plant at Golconda, Nevada, is under construction.
" 1406, line 5 from bottom, *for* alterations, *read* alternations.
" 1417, line 7, *for* Late, *read* Early.
" 1436, Table 1, last column, *for* Saganoga, *read* Saganaga.
" 1463, line 20, *for* other areas, *read* other South shore areas.
" 1498, last line, *for* Lower Huronian, *read* Knife Lake.
" 1503, Table 1, at end, *add* U-unknown.
" 1533, lines 18 and 19, *insert* commas after metamorphism and after authigenic minerals.
" 1741, bottom line, *for* shore, *read* short.
" 1745, at end of paragraph headed Baie Verte, *insert* (Data from K. P. Watson, personal communication)
" 2041, abstract by Harold T. Stearns already printed on page 1948 should not be included here.

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BULLETIN

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	Pages	Plates	Figures	Price to Fellows	Price to Public
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Number 8 (August)	1107–1270	18	32	1.20	1.80
Number 9 (September)	1271–1428	13	36	1.00	1.50
Number 10 (October)	1429–1626	11	28	1.20	1.80
Number 11 (November)	1627–1788	16	33	1.15	1.75
Number 12 (December)					
Part 1	1789–1918	18	25	1.00	1.50
Part 2	1919–2042	—	—	0.60	0.90

Brochures of Volume 51 (Part II)

JULY	Pages	Plates	Figures	Price to Fellows	Price to Public
Elongate drift hills of southern Illinois. By JOHN R. BALL	951– 970	1	11	$0.10	$0.15
Concentric patterns in the granites of the Llano-Burnet region, Texas. By DAVID KEPPEL	971–1000	5	12	0.25	0.40
Experimental flow of rocks under conditions favoring recrystallization. By DAVID GRIGGS	1001-1022	2	10	0.15	0.25
"Flow" in stressed solids: an interpretation. By ROY W. GORANSON	1023–1034	—	—	0.10	0.15
Equilibrium between vapor and liquid phases in the system CO_2-H_2O-K_2O-SiO_2. By G. W. MOREY and MICHAEL FLEISCHER	1035–1058	—	7	0.10	0.15
Differentiation of the Palisade diabase. By FREDERICK WALKER	1059–1106	2	9	0.30	0.45

AUGUST	Pages	Plates	Figures	Price to Fellows	Price to Public
allel folds. By J. B. MERTIE, JR...	1107–1134	—	6	$0.15	$0.25
allel folds. By J. B. MERTIE, JR...	1107–1134	—	6	0.15	0.25
Rock series in diabase sills at Duluth, Minnesota. By GEORGE M. SCHWARTZ and ADOLPH E. SANDBERG	1135–1172	6	11	0.30	0.45
Glaciation in the Wasatch Plateau, Utah. By EDMUND M. SPIEKER and MARLAND P. BILLINGS	1173–1198	5	3	0.25	0.40
Quaternary history of Louisiana. By RICHARD JOEL RUSSELL	1199–1234	—	4	0.20	0.30
Ep-Archean and Ep-Algonkian erosion surfaces, Grand Canyon, Arizona. By ROBERT P. SHARP	1235–1270	7	8	0.30	0.45
SEPTEMBER					
Rock glaciers in the Colorado Front Range. By R. L. IVES	1271–1294	2	5	0.15	0.25
Origin of the muck-silt deposits at Fairbanks, Alaska. By RALPH TUCK	1295–1310	2	3	0.10	0.15
Type locality of the Cretaceous Chico formation. By J. A. TAFF, G. D. HANNA, and C. M. CROSS	1311–1328	2	1	0.15	0.25
Glaciation of the Grays Lake, Illinois, quadrangle. By W. E. POWERS and G. E. EKBLAW	1329–1336	—	6	0.10	0.15
New evidence for an Illinoian glacial boundary in northeastern Ohio. By H. A. IRELAND	1337-1358	1	9	0.15	0.25
Tungsten-bearing manganese deposit at Golconda, Nevada. By P. F. KERR	1359–1390	5	6	0.25	0.40
Gravity section across the Sierra Nevada. By W. D. JOHNSTON, JR.	1391–1396	—	2	0.10	0.15
Time and stratigraphic terminology. By A. H. SUTTON	1397–1412	—	—	0.10	0.15
Structural and volcanic problems in the Southern Absaroka Mountains. By J. T. ROUSE	1413–1428	1	4	0.10	0.15
OCTOBER					
Studies of the Lake Superior pre-Cambrian by accessory-mineral methods. By S. A. TYLER, R. W. MARSDEN, F. F. GROUT, and G. A. THIEL	1429–1538	4	20	0.65	1.00

	Pages	Plates	Figures	Price to Fellows	Price to Public
Igneous rocks of the Terlingua-Solitario region, Texas. By J. T. LONSDALE	1539–1626	7	8	$0.60	$0.90
NOVEMBER					
Pleistocene artifacts and associated fossils from Bee County, Texas. By E. H. SELLARDS	1627–1658	2	7	0.20	0.30
New Pliocene mastodon. By E. H. SELLARDS	1659–1664	—	3	0.10	0.15
Surface of Newfoundland. By W. H. TWENHOFEL and PAUL MACCLINTOCK	1665–1728	5	15	0.40	0.60
Wisconsin glaciation of Newfoundland. By PAUL MACCLINTOCK and W. H. TWENHOFEL	1729–1756	4	6	0.20	0.30
Late Quaternary changes of level in western and southern Newfoundland. By RICHARD FOSTER FLINT	1757–1780	5	2	0.20	0.30
Marine Pleistocene fossils from Newfoundland. By H. G. RICHARDS	1781–1788	—	—	0.10	0.15
DECEMBER					
Audubon-Albion stock, Boulder County, Colorado. By ERNEST E. WAHLSTROM	1789–1820	4	9	0.25	0.40
Geophysical investigations in the emerged and submerged Atlantic Coastal Plain. Part IV. Cape May, New Jersey, Section. By MAURICE EWING, GEORGE P. WOOLLARD, and A. C. VINE	1821–1840	6	3	0.20	0.30
Archean metaconcretions of Thunder Lake, Ontario. By F. J. PETTIJOHN	1841–1850	2	1	0.10	0.15
Glacial chronology of the Southern Rocky Mountains. By LOUIS L. RAY	1851–1918	6	12	0.45	0.70

PROCEEDINGS

General Statement

Beginning with Volume 45, the Proceedings of the annual meeting are published as a separate volume, containing the business of the meeting, the reports of standing committees, and the memorials to deceased members. Volumes 45 to 48 inclusive contain the abstracts of papers presented at meetings. Copies of the Proceedings are sent to Fellows, Correspondents, subscribers, and exchanges, without charge.

Each author of a memorial receives 50 copies of his paper without cost and is permitted to order any additional number at cost. Copies of the volume and of various parts thereof may be obtained from the Secretary at the prices quoted.

Description of the Published Volumes

Volumes	Pages	Plates	Figures	Edition
1933, published 1934	481	11	1	1200
1934, published 1935	477	19	..	1200
1935, published 1936	483	19	1	1200
1936, published 1937	433	16	..	1300
1937, published 1938	388	9	3	1300
1938, published 1939	251	12	2	1300
1939, published 1940	292	14	2	1300

Brochures of 1939 Proceedings

	Pages	Plates	Edition
Presentation of Penrose Medal	78–84	..	100
Certificate of Incorporation and Code of By-Laws, Geological Society of America	111–124	..	125
Officers, Correspondents, and Fellows of the Geological Society of America	125–152	..	125
Memorial to Donald C. Barton. W. E. Pratt	153–166	1	75
Memorial to Arthur P. Coleman. F. D. Adams	167–180	1	75
Memorial to Arthur J. Collier. G. B. Richardson	181–185	1	75
Memorial to William E. Ford. Adolph Knopf	187–193	1	75
Memorial to George H. Girty. J. S. Williams	195–205	1	125
Memorial to Alfred Harker. S. J. Shand	207–209	1	75
Memorial to James R. Macfarlane. G. H. Ashley	211–212	1	75
Memorial to Wendell C. Mansfield. J. B. Reeside, Jr.	213–217	1	125
Memorial to Adolf C. Noé. Carey Croneis	219–227	1	125
Memorial to Giuseppe Stefanini. Julia Gardner	229–239	1	75
Memorial to William A. Tarr. E. B. Branson	241–247	1	75
Memorial to Joseph E. Woodman. E. R. Lilley	249–253	1	75
Proceedings of the thirty-ninth annual meeting of the Cordilleran Section. C. A. Anderson, *Secretary*	255–258	..	50
Proceedings of the thirty-first annual meeting of the Paleontological Society. B. F. Howell, *Secretary*	259–269	..	50
Proceedings of the joint sessions of Section E, American Association for the Advancement of Science, and the Geological Society of America in 1939. H. A. Meyerhoff, *Secretary*	271–281	..	50

Memoirs

General Statement

The Society publishes at irregular intervals a series of books, bound in red buckram. Copies of the Memoirs are sent without cost to Fellows, Correspondents, and exchanges. Extra copies may be obtained from the Secretary at the prices quoted.

Description of the Published Volumes

Memoirs	Pages	Plates	Figures	Price	Edition
No. 1. Stratigraphy of western Newfoundland. CHARLES SCHUCHERT AND C. O. DUNBAR	123	11	8	$2.00	1500
No. 2. Paleozoic plankton of North America. RUDOLF RUEDEMANN	141	26	6	2.00	1500
No. 3. Evolution of the Congo Basin. A. C. VEATCH	183	10	8	1.75	1500
No. 4. Geology in the coastal Atlas of western Algeria. R. VAN V. ANDERSON	450	19	2	3.50	1500
No. 5. Structural behavior of igneous rocks. ROBERT BALK	177	24	38	1.75	1500
No. 6. Structural Petrology. E. B. KNOPF AND EARL INGERSON	270	27	81	2.25	1500
No. 7. Adirondack igneous rocks and their metamorphism. A. F. BUDDINGTON	354	21	30	3.25	1200

SPECIAL PAPERS

General Statement

The Society publishes at irregular intervals a series of Special Papers, bound in paper covers. Copies are sent without cost to Fellows, Correspondents, and exchanges. Extra copies may be obtained from the Secretary at the prices quoted.

Description of the Published Volumes

Special Papers	Pages	Plates	Figures	Price	Edition
No. 1. Bibliographic index of Paleozoic Ostracoda. R. S. BASSLER AND BETTY KELLETT	500	..	24	$2.50	1500
No. 2. Fossil Crustacea of the Atlantic and Gulf Coastal Plain. M. J. RATHBUN	160	26	2	1.25	1500
No. 3. Fossil non-marine Mollusca of North America. JUNIUS HENDERSON	313	..	..	1.75	1500
No. 4. Nuculid bivalves of the genus *Acila*. H. G. SCHENCK	149	18	15	1.25	1000
No. 5. Geology of Saint John, New Brunswick. A. O. HAYES AND B. F. HOWELL	146	9	4	1.25	1500
No. 6. History of the Grand Canyon of the Yellowstone. A. D. HOWARD	159	21	31	1.50	1500
No. 7. Atlantic Submarine Valleys of the United States and the Congo Submarine Valley (with 5 charts). A. C. VEATCH AND PAUL A. SMITH	101	10	28	4.00	1500
No. 8. Middle Devonian corals of Ohio. G. A. STEWART	120	20	2	1.25	1000
No. 9. Fossil Snakes of North America. C. W. GILMORE	96	4	38	.75	1000
No. 10. Stratigraphy and Mollusca of the Eocene of Western Oregon. F. E. TURNER	130	22	7	1.25	1000
No. 11. Geology and paleontology of the Mingan Islands, Quebec. W. H. TWENHOFEL (with A. F. FOERSTE AND M. S. WHITING)	132	24	1	1.25	1000

Special Papers	Pages	Plates	Figures	Price	Edition
No. 12. Tertiary flora from the Rio Pichileufu, Argentina. E. W. BERRY	150	56	11	$1.75	1000
No. 13. Ozarkian and Canadian Brachiopoda. E. O. ULRICH AND G. A. COOPER	324	58	14	3.25	1000
No. 14. Devonian ammonoids of America. A. K. MILLER	262	39	41	2.25	1000
No. 15. Cambrian system (restricted) of the Southern Appalachians. C. E. RESSER	140	16	..	1.25	1000
No. 16. Lower Cretaceous deposits in California and Oregon. F. M. ANDERSON	339	84	3	3.50	1000
No. 17. Pennsylvanian correlations in the Eastern interior and Appalachian coal fields. H. R. WANLESS	130	9	8	1.25	1000
No. 18. Cambrian stratigraphy and trilobites of northwestern Montana. CHARLES DEISS	136	18	7	1.25	1000
No. 19. Cambrian Merostomata. G. O. RAASCH	146	21	14	1.25	1000
No. 20. Geology along the Southern Margin of the Absaroka Range, Wyoming. J. D. LOVE	134	17	3	1.50	1000
No. 21. Paleontology and stratigraphy of the Silurian rocks of the Port Daniel-Black Cape region, Gaspé. S. A. NORTHROP	302	28	1	2.50	1000
No. 22. Lower Mississippian rocks of the east-central Interior. P. B. STOCKDALE	248	26	2	2.00	1000
No. 23. Bibliographic index of North American Devonian Cephalopoda. E. M. KINDLE AND A. K. MILLER	180	..	..	1.00	1000
No. 24. Contributions to a knowledge of the lead and zinc deposits of the Mississippi Valley region. Edited by E. S. BASTIN	156	4	27	1.00	1000
No. 25. Devonian stratigraphy and paleontology of the Roberts Mountains region, Nevada C. W. MERRIAM	114	16	7	1.00	1000
No. 26. Permian ammonoids of the Guadalupe Mountain region and adjacent areas. A. K. MILLER AND W. M. FURNISH	242	44	59	2.25	1000
No. 27. Review of the Pelycosauria. A. S. ROMER AND L. I. PRICE	538	46	71	3.50	1000
No. 28. Bibliography of fossil vertebrates 1928-1933. C. L. CAMP AND V. L. VANDERHOOF	503	..	..	3.50	1000
No. 29. Phacopid trilobites of North America. By DAVID M. DELO	136	13	..	1.00	1000

BIBLIOGRAPHY AND INDEX

General Statement

The Society publishes yearly a series of Bibliographic Contributions, bound in paper covers. The edition is 1800 copies. Copies are sent without cost to Fellows, Correspondents, and exchanges. Extra copies may be obtained from the Secretary at the prices quoted.

Description of the Published Volumes

Volumes	Pages	Price
No. 1. Bibliography and index of geology exclusive of North America. 1933. J. M. NICKLES AND R. B. MILLER	405	$2.00
No. 2. Bibliography and index of geology exclusive of North America. 1934. J. M. NICKLES AND R. B. MILLER	420	1.75
No. 3. Bibliography and index of geology exclusive of North America. 1935. J. M. NICKLES, MARIE SIEGRIST, AND ELEANOR TATGE	408	1.75
No. 4. Bibliography and index of geology exclusive of North America. 1936. J. M. NICKLES, MARIE SIEGRIST, AND ELEANOR TATGE	480	1.75
No. 5. Bibliography and index of geology exclusive of North America. 1937. J. M. NICKLES, MARIE SIEGRIST, AND ELEANOR TATGE	510	2.55
No. 6. Bibliography and index of geology exclusive of North America. 1938. J. M. NICKLES, MARIE SIEGRIST, AND ELEANOR TATGE	471	2.30
No. 7. Bibliography and index of geology exclusive of North America. 1939. J. M. NICKLES, MARIE SIEGRIST, AND ELEANOR TATGE	522	2.60

HISTORY

General Statement

The Society also has on hand a limited number of copies of the book by Herman LeRoy Fairchild, "The Geological Society of America, 1888-1930; a Chapter in Earth Science History," published in 1932 and containing 232 pages and 18 plates. The edition of this book was 1000 copies. Re-issued in 1933, the edition was 300. Copies may be obtained from the Secretary, the cost being $2.50.